CHAPTER – 5

DISTRIBUTION OF PHYSICAL PROPERTIES IN THE OCEANS & IN INDIAN OCEAN

Temperature, salinity, density and oxygen are the three main entities that determine the water characteristics of the oceans. These quantities vary from place to place and from time to time in the oceans and from their distributions we can learn a good deal of about the mean circulation of the waters.

A salient feature of the distributions of many water characteristics is that they are horizontally stratified. In other words, the sea is made up of number of layers as far as these characteristics are concerned, and horizontal changes are much smaller than vertical changes in the same distance. For instance, near the equator the temperature of the water may drop from 25^{o}C at the surface to 5^{o}C at a depth of one kilometer, whereas to get the same temperature of 5^{o}C on the horizontal surface north or south of the equator one should go to about 5000 km which means the average vertical temperature gradient is 5000 times the horizontal one. In addition to spatial variations (vertical and horizontal) there are temporal variations (time variations). This means water properties change with season. Before giving description of these properties, the following statistics gives general information about the ocean character.

a) 75% of the total volume of the ocean water has properties within the range from 0^{o}C to 6^{o}c in temperature and 34 to 35‰ in salinity as shown in Figure 5.1. 75% of the ocean's water has a temperature and salinity within the green region, 99% has a temperature and salinity within the region colored in cyan. The warm water outside the 75% region is confined to the upper 1000 m of the ocean.

b) 50% of the total volume of the oceans has properties between 1.3^o and 3.8^{o}C and between 34.6 and 34.8 psu (Practical Salinity Unit).

c) The mean temperature of the world ocean is 3.5^{o}C (Fig.5.2) and the mean salinity is 34.7 psu.

One noteworthy feature of the spatial distribution of water characteristics is the value of the property is almost same across the ocean in the east-west direction (zonal direction) but may change rapidly in the north - south direction (meridional direction).

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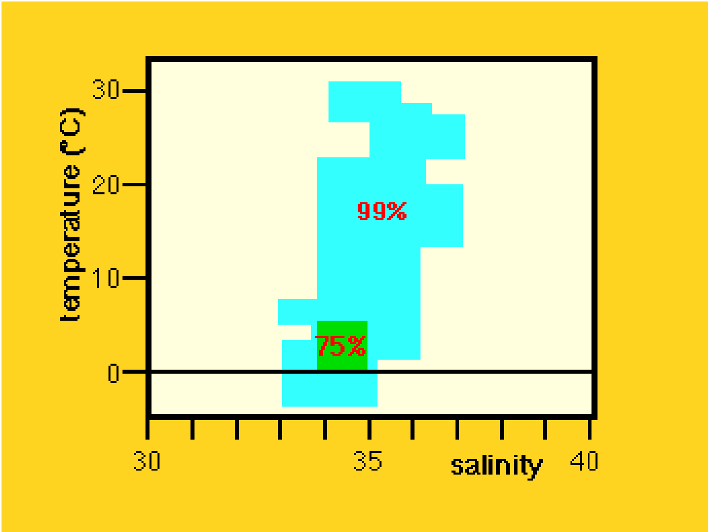


Fig.5.1. Volumetric temperature-salinity diagram of the world ocean

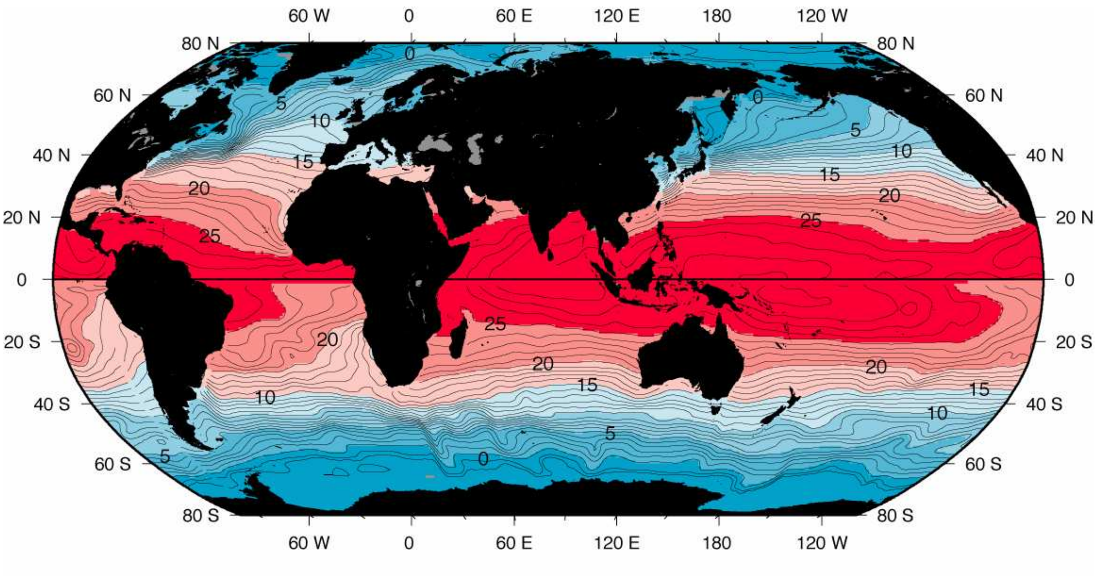


Fig.5.2. Surface temperature: note where the 3.5°C isotherm occurs (most ocean volume is colder than this)

5.1. Density distribution in the Oceans:

The horizontal distribution of density (designated by sigma-t, σ_t, density *in situ* referred at sea surface pressure) in the oceans can be considered as zonal and meridional distributions. While the zonal (east-west) variation of density is almost negligibly small, its meridional variation is considerably large. Water density (σ_t) increases from about 22σ_t near the equator to 26 to 27 σ_t at 50^o to 60^o latitude. (Figure 5.3 shows about the latitudinal variation of temperature, salinity and σ_t at the surface.

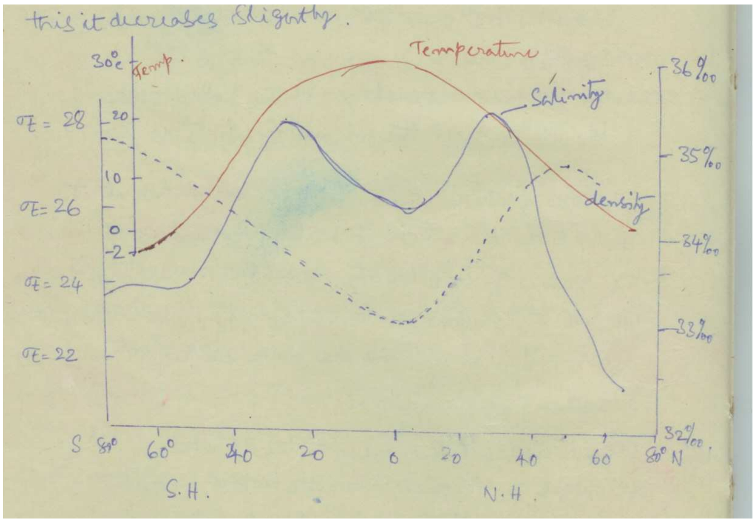


Fig.5.3. Latitudinal variation of temperature, salinity and density

More important, however, is the vertical distribution of density in the oceans. Ocean can be considered into three layers. In the equatorial and tropical regions (30°S-30°N) there is usually a shallow surface layer of nearly uniform density, then a layer of maximum increase and a deep layer of slow increase in density. The first top layer is called as the 'mixed layer', the second one is called as the pycnocline layer and the third one is called as the stratified layer (Fig.5.4).

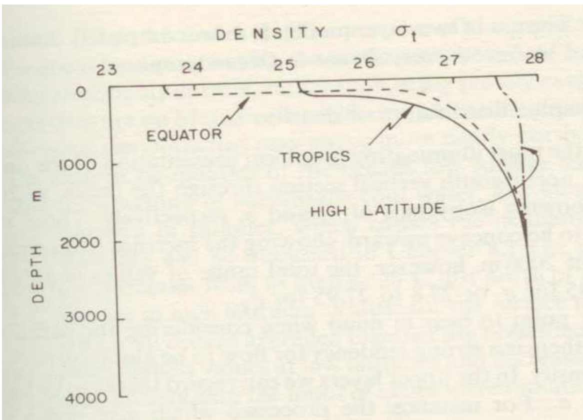


Fig.5.4. Vertical distribution of density in different latitudes

In the Figures 5.5 (a) and (b), if you compare the vertical distribution of density (or the inverse property of σ_t called specific volume anomaly) along the sections A and B, you may note the sharp variation of pycnocline at A than B(Fig 5.5b).

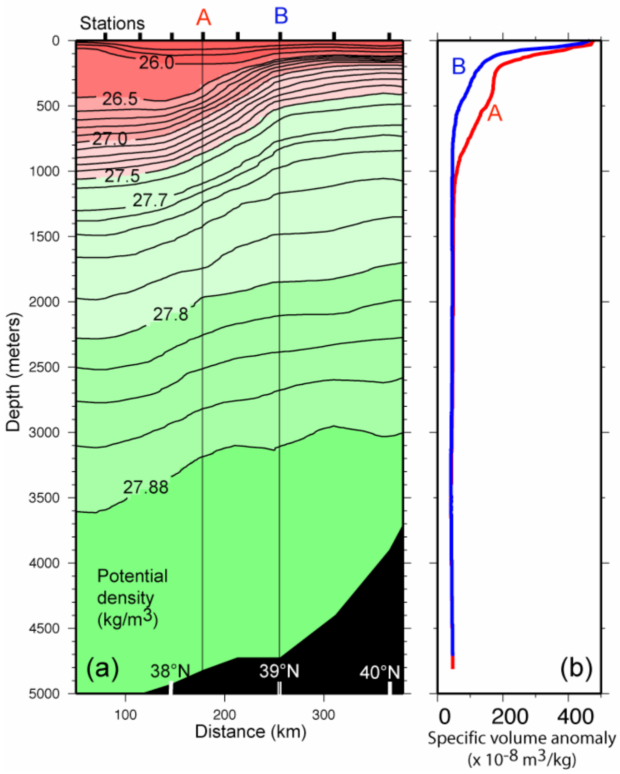


Fig.5.5 comparison of profiles of density at two different sections

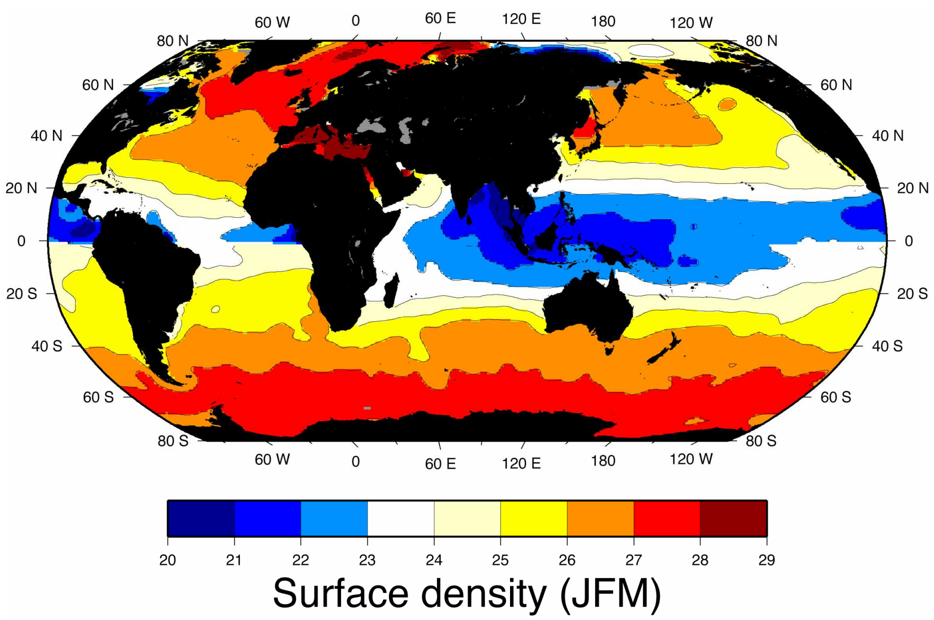


Fig.5.6 Incursion of Low density pacific water into Indian Ocean is seen from density distribution (blue color)

In the Figure 5.6 one may note the horizontal distribution of density at the surface in the global oceans. The royal blue region is between 21-22 σ_t and navy blue is between 22-23 σ_t. This distribution shows that low density water from the Pacific is entering into the Indian Ocean

through the Indonesian region which is called as Indonesian through flow or incursion of Pacific water into Indian Ocean.

The rate of change of the density with depth ($d\sigma_t/dz$ or $d\sigma_\theta/dz$, the potential density gradient) determines the water's stability. If $d\sigma_\theta/dz$ less than, equal to or greater than zero it is classified as unstable, neutral or stable ocean respectively. Where the water stability is high, vertical movement and vertical mixing are minimum, where there is no change of potential density with depth the water is neutrally stable and can mix up easily and is highly unstable if the water stability is low or negative. In fact if a water parcel is lifted, if it goes up little and comes back to its initial position it is called stable, other wise if it swings up and down a little it is called neutrally stable and if it never returns to its initial position it is called unstable.

5.2. The Brunt Vaisala's Frequency:

As the ocean is stratified vertically, the stratification is quantified by the measured value of $\Delta\rho/\Delta z$. The stratification creates a restoring force on the water. If water is displaced vertically, it oscillates in an internal wave with frequency:

$$N = \sqrt{\frac{-g\Delta\rho}{\Delta z}}$$

If the water is more stratified, this frequency is higher. If less stratified, frequency is lower.

For practical calculation of $\Delta\rho/\Delta z$ to get exact frequency, and to get an exact measure of how stable the water column is it can be done like this:

Use a reference pressure for the density in the middle of the depth interval that you are calculating over (for instance, you might have observations every 10 meters, so you would reference your densities at the mid-point of each interval, i.e. change the reference pressure every 10m.

The values of Brunt-Vaisala frequency are 0.2 to 6 cycles per hour. These are also the frequencies of "internal waves". Fig 5.7 shows the comparison of potential density with the Brunt –Vaisala frequency which is a reflection on stability of the waters in the oceans.

Comparing with the frequency of surface waves, which is around 50-500 cycles per hour (1 per minute to 1 per second), internal waves are much slower than surface waves since the internal water interface is much less stratified than the sea-air interface, which provide the restoring force for the waves.

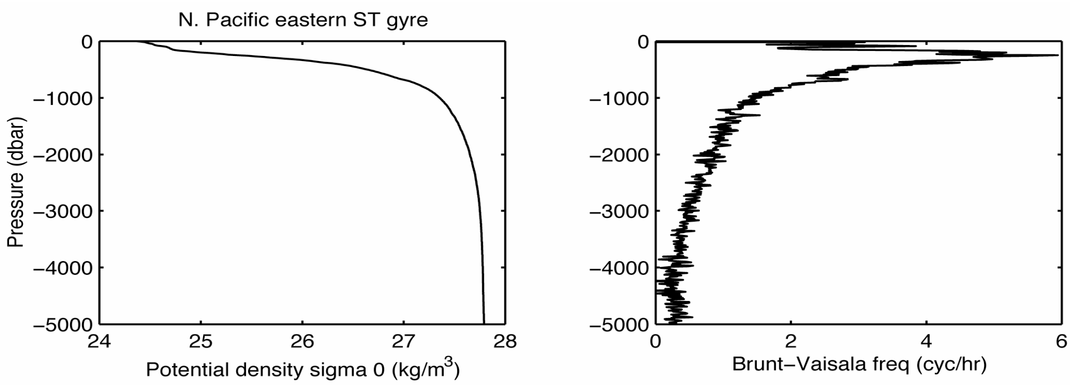


Fig.5.7. Comparison between potential density and Brunt-Vaisala frequency

The water in the depth zone where density increases rapidly with depth is called pycnocline and is very stable. That is to say, it takes much more energy to displace it up or down than in region where density changes slowly with depth. A result is that turbulence, which causes most of the mixing between different water bodies, is unable to penetrate through this stable layer. The pycnocline then, although it is too thin to offer any barrier to the sinking of bodies which are much denser than water, offers a real barrier to the passage of water and water properties in the vertical direction, either up or down.

A further point to bear in mind when considering the ocean circulations is that there is a strong tendency for flow to be along surfaces of constant potential density (σ_θ). Since Deep Ocean water is of high density this implies that it must have been formed at high latitudes because the high density water is found at the surface only at higher latitudes. After formation it sinks along the constant density surfaces and occupies the deep ocean areas of the low latitudes as shown in the Figure 5.8. Thus the Figure 5.8 shows the flow from surface into interior through isopycnals in the oceans. It may be noted in the Figure 5.8, that $\sigma_\theta = 45.85$ (the water density with reference to 4000 db pressure) come from poles to equator. In this Figure 5.8 the blue region indicates the mixed layer and the cyan color zone indicates the pycnocline region while the red zone indicates deep and bottom layers.

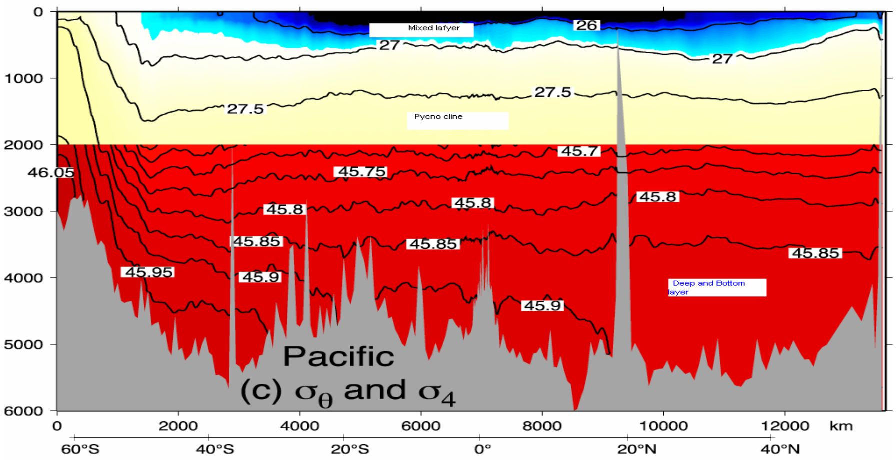


Fig.5. 8 Distribution of density in the Pacific Ocean.

5.3. Density Distribution in the Tropical Indian Ocean (TIO):

In general, surface water density in tropical oceans decreases either by increase of temperature or by decrease of salinity (through precipitation and addition of runoff from river waters) and increases either by decrease of temperature or by increase of salinity (through evaporation). In the northern Indian Ocean, at the Head of the Bay of Bengal the density is low ($< \sigma_t = 20.0$) because of low salinity resulting from heavy monsoon rains and large influx of fresh water from the rivers, while in the northern Arabian Sea, the density is relatively high ($> \sigma_t = 23.0$) because of high salinity resulting from evaporation and advection of waters from Persian Gulf and Red Sea.

In the equatorial and tropical regions, there is a thin layer of low and uniform density near the surface called mixed layer depth (MLD). Estimates of mixed layer depth are important to a variety of investigations including upper ocean productivity, air-sea exchange processes and long term climate change. In the absence of direct turbulent dissipation measurements, the MLD

is commonly derived from oceanic profile data using threshold difference method. The depth at which density (sigma-t) is more than 0.2 kg/m^3 from the surface value is defined as MLD for its estimation.

The depth of mixed layer varies between 10 m and 140 m in the Tropical Indian Ocean (TIO). MLD is lowest (< 20 m) in the Bay of Bengal and Arabian Sea during northern spring (April) and at 10°S-65°E in January. Significant deepening of the mixed layer (50-75 m) was reported in the central Arabian Sea during the onset phase of summer monsoon.

Highest MLDs (>90 m) are found in the central Arabian Sea during the peak summer monsoon (July – August) due to strong wind driven vertical mixing. MLDs are higher (>70 m) during the winter monsoon (January-February) in the central and northern Arabian Sea due to convective vertical mixing. In the northern most extremity of Arabian Sea the mixed layer is particularly deep (>100 m) during January and February. In the equatorial zone, MLDs generally vary from 20-30 m in April to 50-60 m in July. A region of high MLD (>70 m) is located in the Southeastern Indian Ocean (around 15°S- 100°E) during September. Between 20° S and 30°S, MLDs are highest (>100 m) during southern winter (August-September) and lowest (<40 m) during southern summer (December - January).

The heat content of the water column within the MLD is very important for the synoptic weather disturbances. MLD and the Mixed Layer Heat Content (MLHC) were found to be more correlated to buoyancy flux and net heat flux in the central western equatorial Indian Ocean during pre-onset and onset phases of summer monsoon.

Using the monthly mean temperature data over 1°×1° grid for standard depths (0, 10, 20, 30, 40,50, 75, 100, 125 and 150 m) for the TIO that are extracted from any atlas like World Ocean Atlas (WOA, 2001 and 2005), the monthly mean MLDs can be estimated from the density difference criteria,(as explained in the above section), and MLHC can be computed from the equation:

$$MLHC = \rho C_P \int_0^D \bar{T}\, dz$$

where ρ is *in situ* density of sea water, C_p is specific heat of sea water at constant pressure, $\bar{T}$ is the mean temperature of the layer thickness, dz. D is the Mixed Layer Depth (MLD).

Heat content of the mixed layer varies between 10×10^8 and 110×10^8 J/m ^{2} in the TIO. MLHC generally followed the pattern of MLD variation. MLHC is low ($< 40\times 10^8$) in the region north of 10° S during March-May. It is high ($> 70\times 10^8$ J/m ^{2}) in the northern and western Arabian Sea during January- February and highest ($> 80\times 10^8$ J/m ^{2}) in the south central Arabian Sea during July-August. It is lowest ($< 20\times 10^8$ J/m ^{2}) in the northern Bay of Bengal during August-September. In the equatorial Indian Ocean (along 5° S) MLHC is lowest ($< 20\times 10^8$ J/m ^{2}) during December-January (southern summer) and highest ($> 50\times 10^8$ J/m ^{2}) in the eastern part during August. In the southern TIO (10° S - 30° S) the MLHC is highest ($> 90\times 10^8$ J/m ^{2}) during July-September and lowest ($< 30\times 10^8$ J/m ^{2}) during December-January.

Just beneath the MLD, there lies the pycnocline and below this, density increases slowly towards the deeper layers. In the north Indian Ocean the depth of pycnocline exhibits spatial and seasonal variations. In the north equatorial part during the northeast monsoon period along the equator (east of 76° 40' E) and from equator to 3° N in the eastern part the pycnocline is sharply developed with steep density gradients and starts (along with thermocline) at 95 to 100 m depth. North of 3° N the pycnocline gradually rises towards the surface and of the east coast of Sri Lanka (along 83° E) it is seen at 50 m depth. Along the equator there is shallowing of pycnocline towards west (from about 95 to 100 m depth in the eastern part to about 30 m depth in the western part) during the monsoon transitions. During the transition period between winter and summer in the western Bay of Bengal (west of 90° E), the depth of pycnocline slightly increases towards south and varies from 75 m to 100 m. In the eastern Bay and the Andaman Sea, it starts normally

at comparatively lower depth. The low salinity of surface waters of the Bay of Bengal causes in all seasons to be isolated from the deep waters by a sharp pycnocline between 50 and 100 m. The seasonal variation of surface density is resulted due to insolation, precipitation, evaporative cooling and influx of saline and fresh water. During the pre-summer monsoon season low density ($<\sigma_t = 20.5$) waters occupy the east central Bay near the Andaman Islands and relatively dense (σ_t=21.0) waters are found along the coast and offshore areas. During the summer monsoon, the lowest density ($<\sigma_t = 13$) waters are seen in the north western Bay and the distribution pattern of density in this area resembles that of the salinity. Density increases gradually towards the south and tongue of low density water spreads south-westwards in the central Bay. South of 12° N density increases to $\sigma_t \geq 22.0$ in the southern Bay where relatively high salinity waters are noticed. During the post-summer monsoon season, surface density varies between σ_t=20.0 and σ_t=22.0 and a flow of dense waters, characterized by high salinity, from the southwestern Bay is seen up to 8° N. During the winter monsoon, density of the surface waters is less than σ_t=21.8. Two cells of very low density are found off the central east coast of India and northwest of the Andaman Islands, where fresh water discharges enters the Bay of Bengal from the Krishna and Irrawady rivers respectively. A large tongue of high density water, protruding northwards, appears in the central part of the Bay between 5° N and 15° N and separates the two cells of low density water. Spreading of low density water from the Andaman Sea into the Bay of Bengal through the Ten Degree channel is evident. During the southwest monsoon a tongue of high density water with its axis oriented in a SW-NE direction is noticed in the central Bay at depths of 50 m and 100m. At 100 m depth during the pre-summer monsoon season a large cell of low density water with core value of σ_t=22.0 is seen between 12° N and 17° N. During the summer monsoon season, a broad band of high density water ($>\sigma_t$=24.5) is seen in the central Bay. During the post -summer monsoon season, density varies between σ_t=24.0 and σ_t=25.0 in the southern Bay of Bengal. During the winter monsoon season density varies between σ_t=24.96 and σ_t=21.8. During the pre-monsoon season and summer monsoon season the low density cells observed at 100 m depth persists upto 200 m depth also. In the western Bay the distribution is characterized with alternate cells of low (σ_t=26.0) and high (σ_t=26.4) dense waters. During the post-summer monsoon season density varies between σ_t=25.8 and σ_t=26.4. During the winter monsoon season density is low in the north western Bay and relatively high off the central east coast of India and the coast of Burma. At 500 m depth two cells of high density appear in the northern and western Bay during the pre-summer monsoon season. High density waters are noticed in small cells during summer monsoon season also.

5.4. DISTRIBUTION OF TEMPERATURE:

5.4.1. HORIZONTAL DISTRIBUTION:

As said earlier, in the horizontal distribution of temperature, zonal variations are much lesser than that of meridional variations. Though the isotherms, in general, run parallel to the latitudes (Figs.5.9a & 5.9b), very near the continental boundaries, some sharp kinks are observed. These sharp kinks are due to land and sea variations (climate variations) and in some places due to advection of currents (either warm or cold currents entering into the area). Also due to upwelling in the eastern shores of the oceans these sharp variations are caused.

The meridional variations in temperature are due to differential heating. Though the maximum temperature occurs in the equatorial and tropical regions, the upwelled cold subsurface water in the equatorial region lowers the temperature.

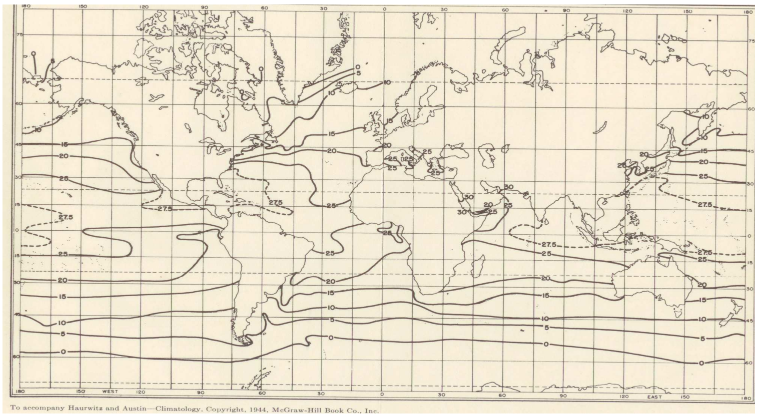


Fig.5.9a. Temperature distribution in the Oceans in August

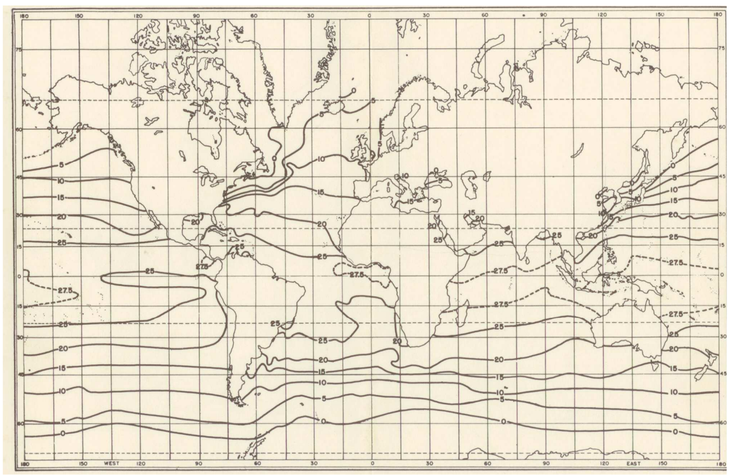


Fig.5. 9bTemperature distribution in the oceans in February

Because the Northern Hemisphere is dominated by land and the Southern Hemisphere by sea, temperature differences between summer and winter are more extreme in the Northern Hemisphere (the land warms and cools more quickly than the ocean). In the Fig.5.10, average land and sea temperatures are shown for each place on Earth, from the North Pole (left) to the South Pole (right). New Zealand is located around 40º S, in the temperate climate zone, where land temperatures change about as much as sea temperatures. At the antipodes on the other side of the globe, the situation is much different. It experiences hotter summers and colder winters. The temperate regions experience the largest seawater temperature changes and New Zealand is no exception. Notice that the difference in temperature from winter to summer, about 6-8ºC, amounts to shifting by 10-12ºC North-South for summer-winter. Note also that polar sea water has practically no temperature change, due to the stabilizing effect of the ice.

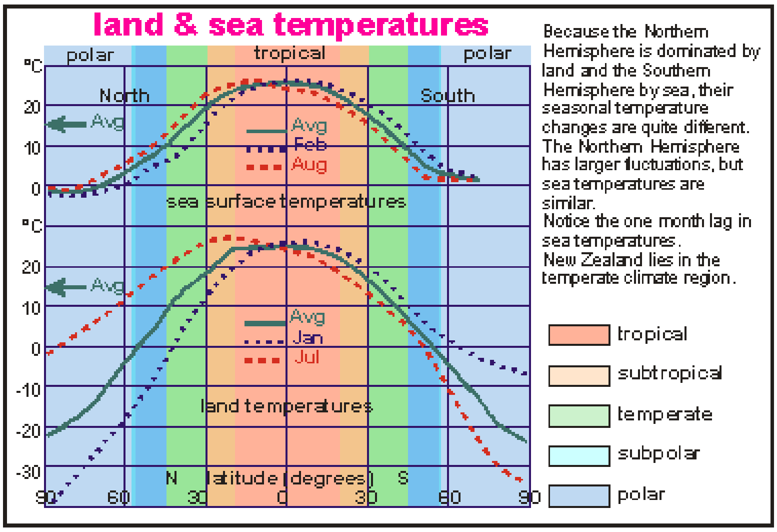


Fig.5.10.Variation of SST in different latitudinal zones

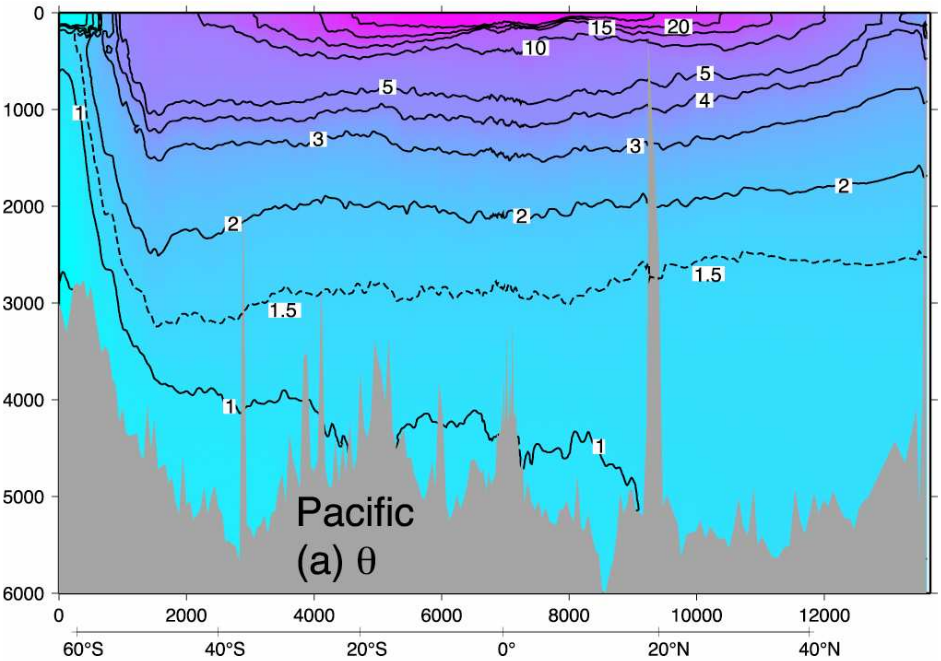


Fig.5.11. Potential Temperature distribution in Pacific Ocean

Figure 5.11 depicts the three dimensional distribution of temperature in the Pacific Ocean. The crimson red colored zone is thermocline layer, deep blue color zone is the upper layer of permanent thermocline, light blue color is the deep zone and green color is the colder bottom zone. While the 15^{0}C isotherm demarcates the main thermocline zone, 3^{0}C demarcates the upper layers of the oceans and 1.5^{0}C demarcates the deep and bottom layers of the ocean. The top

corners on either side of the poles at the surface between 1° and 2°C indicate the dicothermal layer or inversion layer.

5.4.2. Vertical distribution of temperature:

Defant described the ocean temperatures in the vertical, in analogy to the atmosphere, as oceanic troposphere and stratosphere. The only difference between the atmosphere and the oceans is the profiles are reversed and tropopause begins first at the sea surface in low latitudes. This two fold sub divisions of the thermal structure of the ocean is limited to the tropical and subtropical oceans. In the Polar regions the oceanic troposphere becomes less well developed and the oceanic stratosphere comes close to the sea surface. While the average ocean temperature is 3.8°C, at the equator the average temperature is 4.9°C.

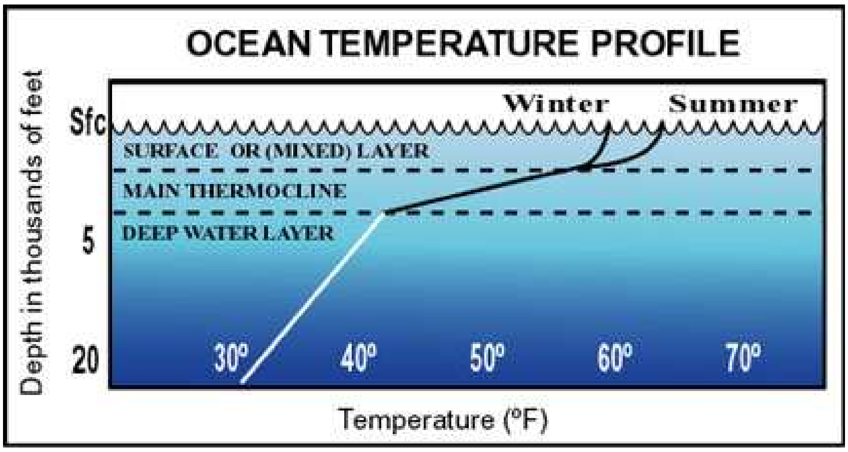


Fig. 5.12 a) Ocean temperature profile

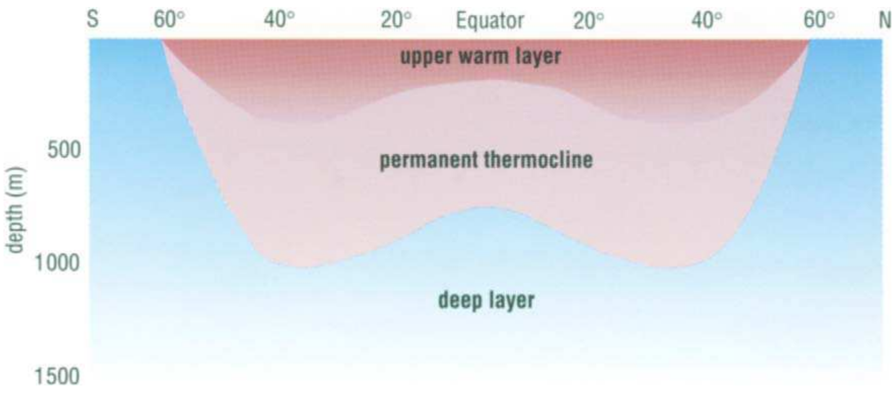


Fig.5.12 b) three dimensional view of the structure of the oceans

The ocean can be divided into three vertical zones as shown in the Figures 5.12 a & b, depending on temperature. The top layer is the **surface layer**, or **mixed layer**. This layer is very easily influenced by solar energy, wind and rain. The next layer is **thermocline**. Here the

temperature changes rapidly with depth, which is found in the temperature range 8 - 15°C, is called the permanent thermocline. It is located at 150 - 400 m depth in the tropics and at 400 - 1000 m depth in the subtropics. The last layer is the **deep-water layer**. Water temperature in this zone decreases slowly as depth increases. Water temperature in the deepest parts of the ocean is about 2°C.

MIXED LAYER DEPTH (MLD):

In the top layer (i.e. 50 to 200 m depth layer), the vertical differences in temperature are negligible. This layer is called ‘surface layer’ or ‘isothermal layer’ or ‘mixed layer’. These isothermal conditions are attained in this layer by mixing of surface waters due to external and internal forces like winds, waves, tides and currents.

Oceanic surface layer is a region of intense mixing and hence the properties within this layer are nearly homogeneous. This nearly homogeneous upper ocean layer is referred to as the mixed layer. The thickness of mixed layer is an important parameter in determining the quantity of heat that is available for exchange with atmosphere which is capable of triggering several ocean - atmosphere coupled processes such as convection, generation of tropical cyclones, etc. Atmospheric forcing that regulates mixing in the upper ocean are winds, waves, solar heating, evaporation and precipitation. Figure 5.13 gives a schematic picture of various factors affecting the mixed layer. The solar heating and precipitation stratify and stabilize the upper ocean, while the wind, wave action and evaporation destabilize it through mixing. Since the atmospheric forcing is highly variable on space-time scales, the geographical location, to a great extent, decides the structure and variability of the upper mixed layer. The MLD is an important parameter in determining the chlorophyll biomass and biological productivity of the upper ocean. The MLD varies on several temporal scales ranging from diurnal, intra-seasonal, seasonal to inter-annual and the variability of MLD is directly linked to processes occurring in the mixed Layer.

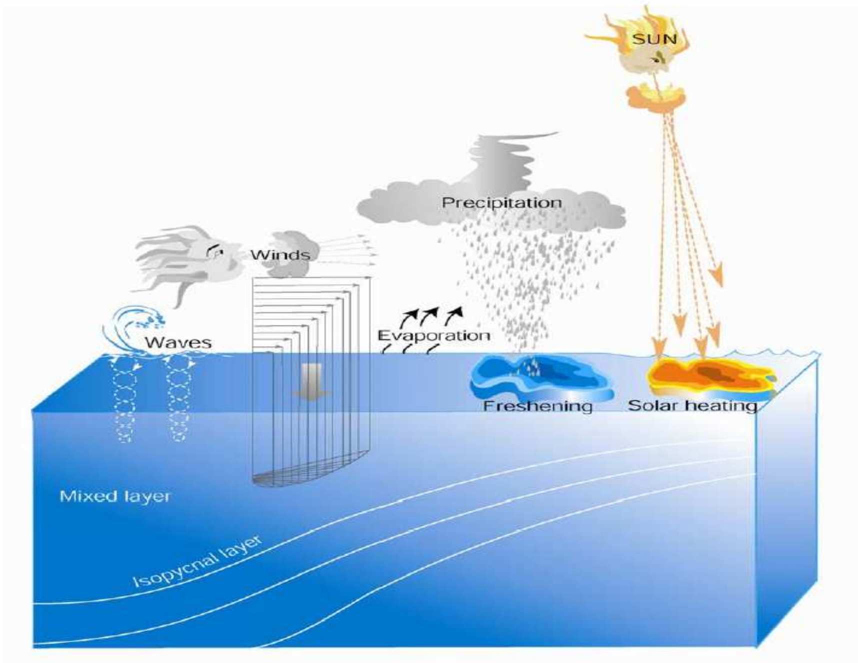


Fig.5.13. Schematic representation of various factors that influence the upper ocean mixed layer (By courtesy of Prasanna Kumar and Narvekar, 2005)

Two types of MLD definitions have been most commonly used in the literature – (1) based on specifying a difference in temperature or density from the surface value (Wyrtki, 1964; Levitus, 1982; Schneider and Muller, 1990) and, (2) based on specifying a gradient in temperature or density (Bathen, 1972; Lukas and Lindstrom, 1991). It is important to examine the distribution of properties within the upper layer before any such criteria are applied.

For example, the vertical profiles of temperature, salinity and sigma-t for the Bay of Bengal are presented for a 1-degree grid centered at 9°N and 19°N latitude along 89°E for the month of August and February (Fig.5.14), which represents the summer and winter conditions respectively.

The vertical profiles indicate that the isothermal, isohaline and isopycnal layers, in general, coincide in the upper ocean irrespective of the season in the southern part of the Bay of Bengal (Fig.5.14 left panels), but in the northern Bay profiles of temperature and salinity show a different vertical structure (Fig.5.14 right panels). In the northern Bay, the temperature shows an isothermal layer within which the salinity increases rapidly with depth. This is associated with freshening due to the river runoff as well as precipitation during summer. Hence, the criteria for defining the MLD whould have to take into account the density variation rather than temperature or salinity. Therefore the authors defined the MLD as the depth at which the density (sigma-t) exceeds 0.2 kg m^{-3} from its surface value. However, the Isothermal layer here is defined as the depth at which the temperature decreases by 1°C from its surface value (Fig.5.14).

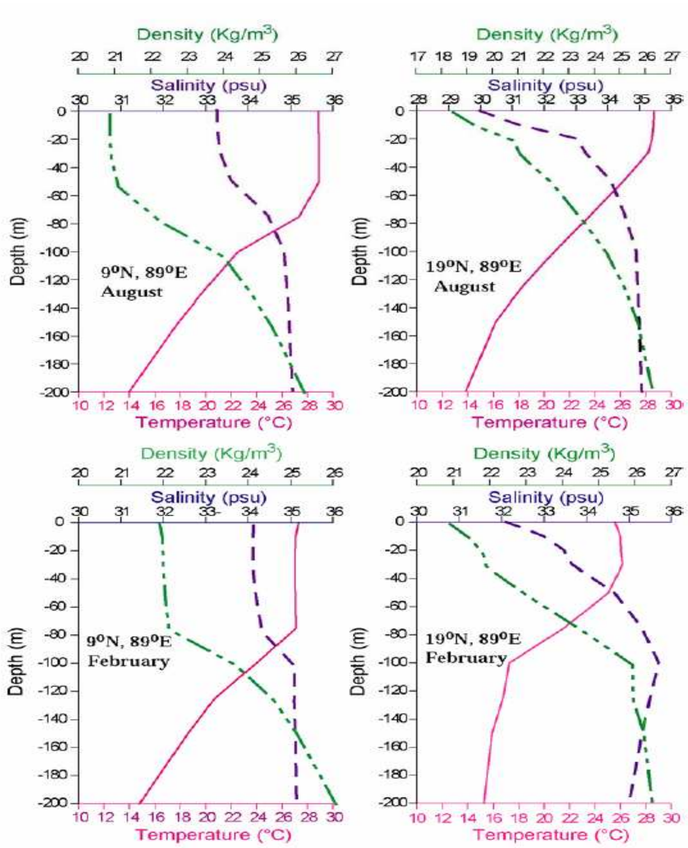


Fig.5.14. Vertical profiles of temperature (pink solid line), salinity (purple dashed line), and density (green dash dotted line) for August (top) and February (bottom) at 9°N (left) and 19°N(right) along 89°E in the Bay of Bengal.(By courtesy of Narvekar, 2008)

4.6 Factors controlling the mixed layer variability:

The factors that are responsible for the observed variability in the mixed layer are found to be heat flux, momentum flux (wind-stress curl) and fresh water flux (evaporation minus precipitation minus runoff).

The MLD during the spring intermonsoon (March-April-May) is thin in the Bay of Bengal compared to the rest of the months. It varies between 10 and 25 m except in the southwestern region in March and April and near the western boundary in April (Fig.5.15a).

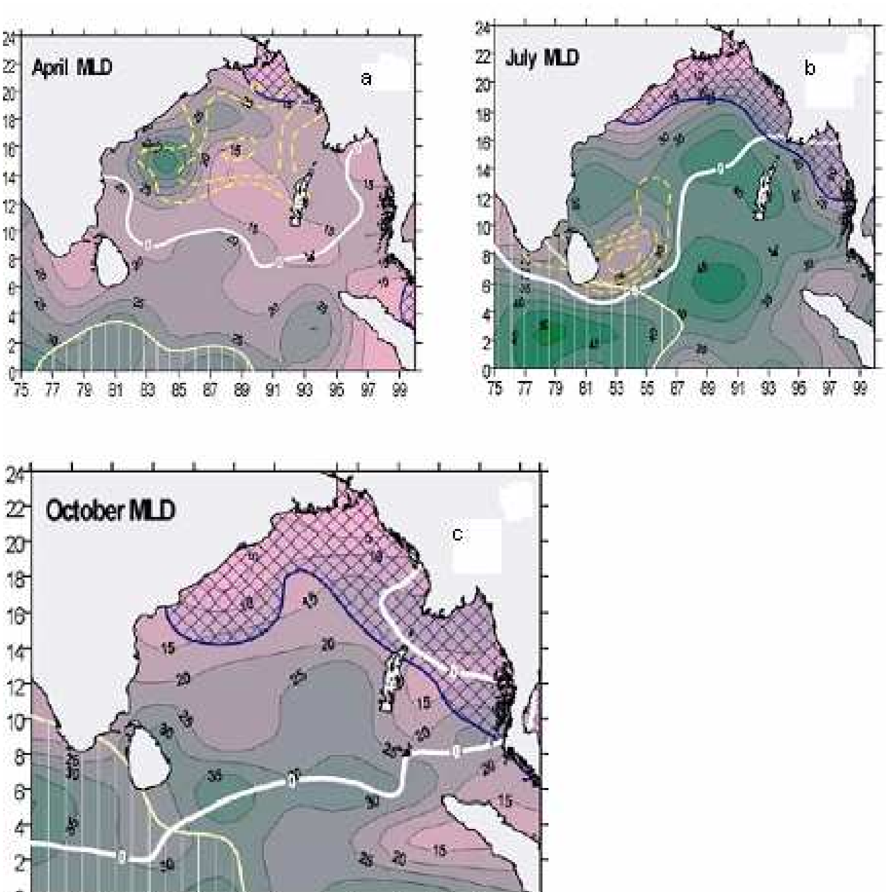


Fig. 5.15. Spatial distribution of mixed layer depth in different seasons. (a) spring intermonsoon, (b) Southwest monsoon and (c) fall intermonsoon. The blue cross-hatch represents the region where salinity is less than 32 psu while vertical lines within the thin yellow solid line represent the region where the salinity is greater than 34.5 psu. The thick broad white line indicates the zero wind stress curl. The yellow broken lines in (b) indicate the negative wind stress curl (-10 to -20 × 10^{-8} Pascal/m) with increasing magnitude towards the center.

Mixed layer during summer monsoon (July) is deep in the zonal band between equator and 6^{o}N (Fig.5.15b) and closer to the western boundary towards the end of spring intermonsoon

(Fig.5.15a). The northern and eastern part of the Bay has shallow MLD. With the progress of summer monsoon, the region of deep MLD expands towards the central and shallow towards northern Bay. The shallow MLD in the northern part of the Bay is due to river discharge and monsoon rainfall while the deep MLD in south is due to intrusion of high salinity Arabian Sea water through Monsoon current.

As the summer monsoon retreats over the Bay of Bengal and the fall intermonsoon sets in, the shallow MLD which was confined to northern Bay, north of 18^{o}N, was seen extending southward to15^{o}N in October (Fig.5.15c).

THERMOCLINE:

It lies underneath the isothermal layer. In this layer temperature decreases to the maximum or the depth at which temperature gradient is highest is called the thermocline. The reason for the existence of thermocline is in low and middle latitudes the temperature of the upper layers is warm due to warm atmosphere above, while the temperatures in the deep layers are very low due to polar origin, as they flow towards low latitudes along the σ_θ surfaces as shown in Fig.5.8. As they flow towards low latitudes through subsurface layers from high latitudes and mix with warmer waters of the upper layers from below gain a little heat, but are continually renewed from high latitudes and are thus kept at a relatively low temperatures while the top layers are continuously kept at a higher temperature due to warm low latitude atmosphere above. Thus due to the difference between high temperatures at the top and low temperatures in the deep layers make the middle layer intermediary. This intermediary layer thus forms a thermocline as a maximum drop in temperature gradient.

In low latitudes there is a thermocline present in the depth range between 200 and 500 m all the time as shown in Fig.5.12b & Fig. 5.16b. This is referred to as 'the main' or permanent thermocline. In polar waters there is no permanent thermocline. But in middle latitudes (Fig. 5.16a) the temperature in the upper zone shows seasonal variations and goes to as much as 1000 m. In winter the surface temperature is low and the mixed layer extends deep. In summer the surface temperature rises and a seasonal thermocline often develops in the upper layers.

At high latitudes (Fig. 5.16c), the surface temperatures are much lower than at lower latitudes, while the deep water temperatures are a little warmer. As a result the main thermocline is present and only a seasonal thermocline may be present. However, in winter due to cooling of the surface layer, the temperature decreases up to 4^{o} C as a result vertical convection takes place. Thus thorough mixing occurs in the surface layers. With further cooling ice forms at the surface layer and so salinity increases in the remaining waters. In summer the upper layers gets heated up and so ice melts up to 30-50 m depth. This is the situation in high latitudes. Beyond the high latitudes in the polar latitudes there is often a 'dicothermal layer' at 50-100 m is formed (Fig.5.16d). This is a layer of cold water, down to -1.6^{o}C, sandwiched between two warm water layers above and below.Some times inverse dicothermal layer with warm water sand wiched between two cold waters above and below is also possible as shown in Fig.5.16c.

If the surface temperature is very low, convection from cooling can reach deeper than the surface layer. This situation is encountered in the Polar Regions where cold water sinks to the bottom of the ocean. This process replenishes the deeper waters and is responsible for the currents below the upper kilometer of the ocean. Areas of deep winter convection are the Weddell Sea and the Ross Sea in the Southern Ocean and the Greenland Sea and the Labrador Sea in the Arctic region.

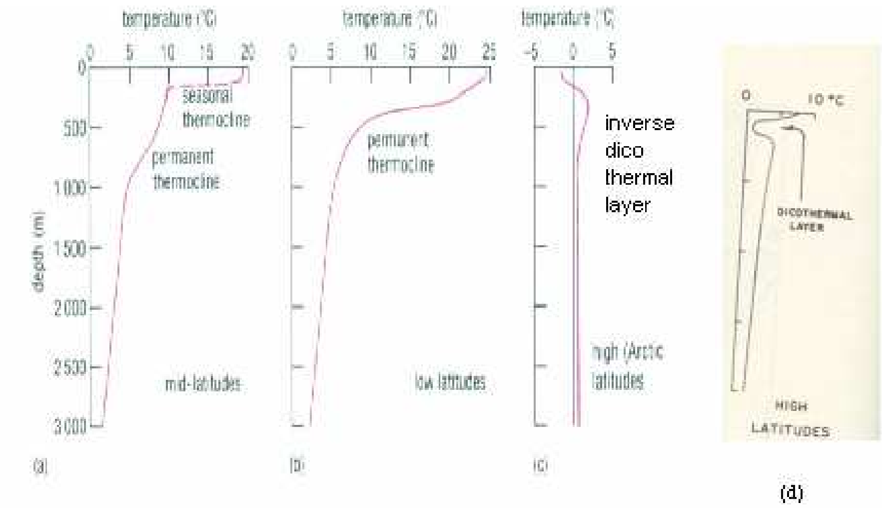


Fig. 5.16. Vertical temperature profiles in different latitudes

The seasonal variation of thermocline (growth and decay) is shown in Fig.5.17. From March to August the temperature gradually increases due to absorption of solar energy so the thermocline rises upward from 30 m in May to 20 m in August and as winter conditions starts, it again goes down from September to January up to 100 m or so.

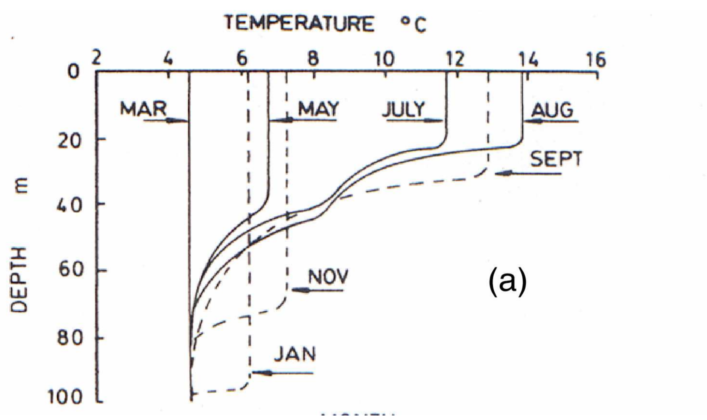


Fig.5.17. A schematic of seasonal variation of thermocline

Compared to air, water has an extremely high heat capacity, so it takes much more sunlight to warm up. Fortunately, warm sea water is lighter than cool sea water, so the warmed water stays on top and is reluctant to pass its heat downward. As a result, the sea warms slowly but cools more quickly. During the summers that is why a thermocline develops between the warm surface water and the cooler bottom water. As the sea warms further, this sharp boundary moves deeper. In tropical and temperate latitudes, the thermocline may descend to 15m. Sometimes a second thermocline is found at 40m depth, originating from the continental shelf. Towards winter, as the surface water cools,

the thermoclines disappear. Also during heavy storms, the sea water may get mixed so thoroughly, that thermoclines disappear.

5.5. Vertical distribution of temperature and salinity in the north Indian Ocean:

The typical profiles of Temperature and Salinity at selected locations of the tropical Indian Ocean are presented Fig.5.18. The temperature variation of the seasonal thermocline in the top 200m is nearly the same at these locations. However, large variations in salinity are

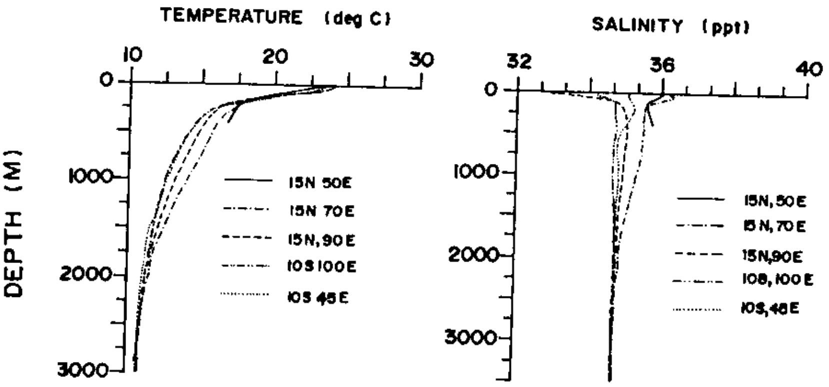


Fig.5.18 Vertical profiles of: (a) temperature and (b) salinity at selected locations in the north Indian Ocean

noticed in the depth range of seasonal thermocline. The salinity profiles at 15°N, 70°E and 10°S, 100°E represent the extreme conditions. The higher salinities at 15°N, 70°E are due to excessive evaporation over precipitation in the Arabian Sea while the lower salinities at 10°S, 100°E are due to excessive precipitation in the southern Indian Ocean and advection of less saline Indonesian through- flow waters. Sharp halocline is present at around 100 m in all the station profiles. Within the halocline, a shallow (or 75m) salinity maximum (>36.3‰) of the Arabian Sea High Salinity Water (ASHSW) is present at 15°N, 70°E and lowest salinities (<34.6‰) are seen at 10°S, 100°E. In the Bay of Bengal, salinity gradually increases to a maximum at around 250m. Off the Arabian Coast, subsurface high salinity (35.5-35.8‰) waters are present in the depth interval 200-400 m. Below the seasonal thermocline, the temperature and salinity variations are large corresponding to the flow of cold, less saline waters from the higher latitudes and warm high saline waters from the lower latitudes. The main thermocline in the Bay of Bengal and the south Indian Ocean is relatively cooler and characterized by lower salinities. Whereas in the Arabian Sea, warm higher salinities characterize the main thermocline due to the presence of the Persian Gulf Water (PGW) and Red Sea Water (RSW) in the depth range of 200-1500m. Maximum variation in T and S at 750m is due to the presence of RSW. In the Bay of Bengal, the salinity profile indicates the admixture of the PGW and RSW. The salinity and temperature profiles at 10°S, 100°E and 10°S, 45°E show the presence of cooler, low salinity waters of the Indonesian through-flow from the Pacific Ocean and their westward advection up to the east coast of Africa, through the South Equatorial Current (SEC). However, the relatively higher salinities at 10°S, 45°E around 250m show the influence of high salinity waters from the Arabian Sea up to the western tropical Indian Ocean.

The vertical distribution of temperature in the Bay of Bengal across a section along 88°E running from 4°N to 20.5°N during July-October is shown in the Fig.5.19 The isotherms in the upper 100m slope upwards towards north, forming a ridge around 19°N. The isotherms also

form a dome like structure between 18°N and 12°N. Further south the isotherms are nearly horizontal.

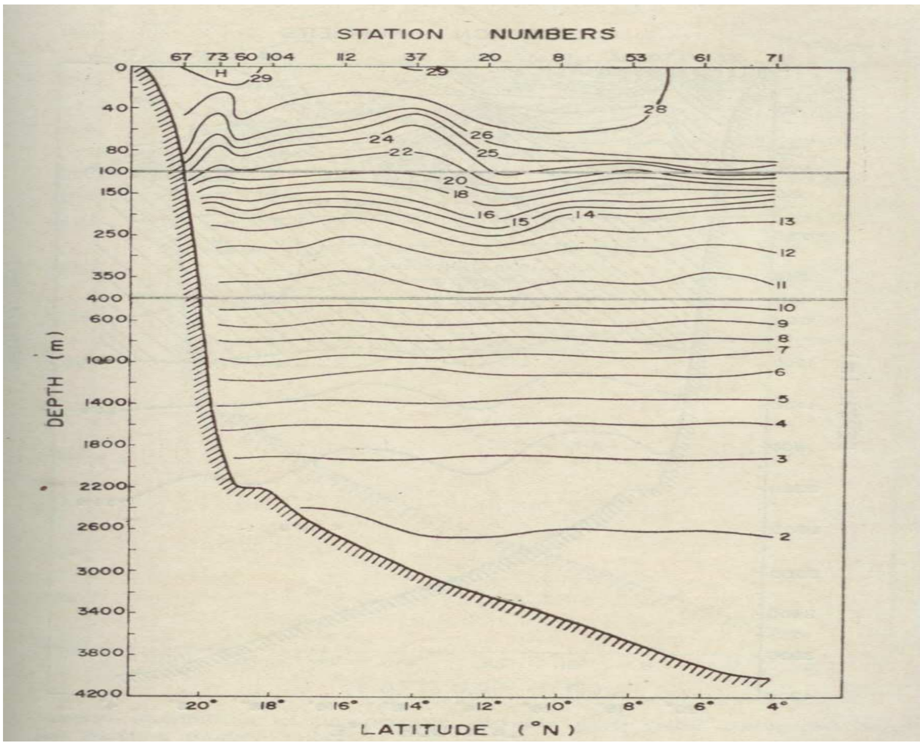


Fig.5.19. Vertical distribution of temperature in the Bay of Bengal. By courtesy of J.S..Sastry et al.(1985) Mahasagar, 18(2),153-162

5.5.1. Temperature Inversion

Surface layer temperature inversion occurs in the South-Eastern Arabian Sea (SEAS) during winter. It is found that the inversion in this area is a stable seasonal feature and the occurrence is limited to the coastal waters. The thickness of inversion layer varies from 10 to 80 m and the temperature difference varies from 0.0 to 1.2° C from surface temperature and the temperature at inversion depth. The causative factor for the inversion is identified to be the lying of winter time surface advection of cold less saline Bay of Bengal water over the warm saline Arabian Sea water along the west coast of India.

The low salinity waters and the inversions propagate westward along with the downwelling Rossby waves that constitute the Lakshadweep sea level high; inversions occur in the western Lakshadweep Sea (LS) (~73° E) about 40 days after they occur near the coast in the eastern LS (75.5 E). They disappear in April when the Tropical Convergence Zone (TCZ) moves over SEAS and the warm pool engulfs the region. Ocean dynamics and air-sea fluxes are together responsible for the formation and westward propagation of the inversions.

The spatially organized temperature inversions occur in the coastal waters of the western and northeastern Bay of Bengal during winter (November-February). Although the inversion in the northeastern Bay is sustained until February (with remnants seen even in March), in the western Bay it becomes less organized in January and almost disappears by February. Inversion is confined to the fresh water induced seasonal halocline of the surface layer. Inversions of large temperature difference (1.6° – 2.4° C) and thin layer thickness (10-20m) are located adjacent to

major fresh water inputs from the Ganges, Brahmaputra, Irrawaddy, Krishna and Godavari rivers. Inter-annual variability of the inversion is significantly high and is caused by inter-annual variability of fresh water flux and surface cooling in the Northern Bay. Fresh water flux leads to the occurrence process in association with surface heat flux and advection. In the western Bay, the East Indian Coastal Current (EICC) brings less saline and cold water from the Head of the Bay to the Southwest Bay, where it advects on to the warm saline water, and promoting temperature inversion in this region in association with surface heat loss. For inversion occurring in the Northeastern Bay (where the surface water gains heat from atmosphere), surface advection of the less saline, cold water from the Head of the Bay and Irrawaddy basin is found to be the major causative factor.

5.5.2. Vertical distribution of temperature and salinity in Arabian Sea and Bay of Bengal:

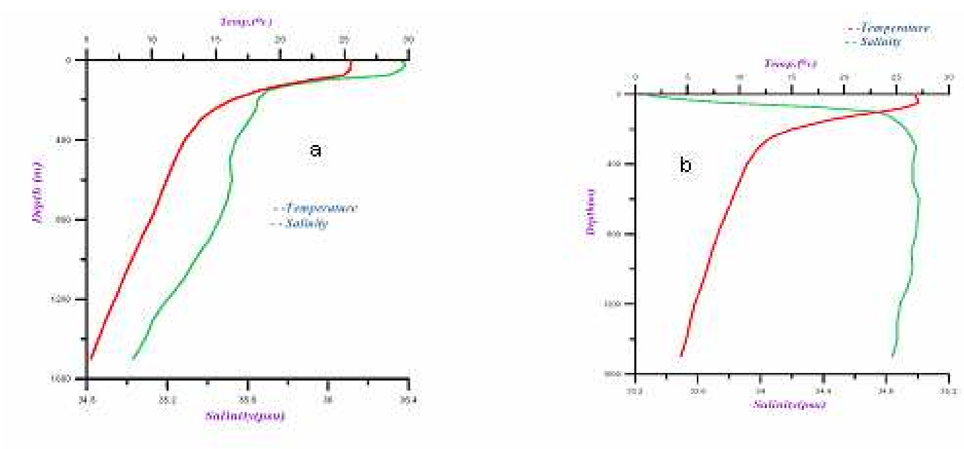


Fig.5.20. Vertical profile of temperature (Red) and salinity (Green) in (a) the Arabian Sea at 14.5^{O}N, 63. 5^{O}E and (b) Bay of Bengal at 14. 5^{O}N, 87. 5^{O}E (Advilkar, 2008)

The vertical profile of annual mean temperature at a location in the Arabian Sea shows an isothermal layer in the upper 50 m and below this the temperature decreases rapidly with depth up to 300 m (thermocline) (Fig.5.20 a). Below 300 m the temperature decreases slowly with depth (stratified layer). The vertical profile of annual mean salinity at the same location (Fig.5.20 a) shows a similar pattern like that of temperature in the upper 40 m (Fig.5.20a). Below 40 m and up to 300 m the salinity decreased rapidly, known as the halocline, while below 300 m the decrease of salinity with depth is slower.

In the Bay of Bengal the thermal structure shows an isothermal layer in the upper 60 m within which the salinity shows a rapid increase (Fig.5.20b). This layer is known as the barrier layer. Below 60 m the temperature decreases rapidly with depth up to 300 m but the salinity by and large remains constant. Below 300 m temperature shows very slow decrease.

5.5.3. Horizontal distribution of temperature in Indian Ocean:

The surface temperatures in Tropical Indian Ocean (TIO) vary from less than 20° C (in the southern part during July – October) to more than 30° C (in the northern part during April – May). In the Persian Gulf and Red Sea the surface temperatures exceed 32° C during July –

August. The distribution of isotherms is mainly zonal south of 10° S. At 10° S the SST varies from about 29° C (in March – April) to about 26° C (in August). At 30° S it varies by about 5° C (22-27° C) in the west and about 2° C (20-22° C) in the east.

In the region north of 10° S the monsoon winds and the associated currents significantly influence the SST. During January-February, between 10° S and 10° N, SST increases from about 28° C in the west to more than 29° C in the eastern part. In February, the thermal equator runs from 10° S off the African coast to the equator near 90° E. During April-May, it moves towards 10° N. It descends southward and is located just south of the equator in July. The equatorial SST minimum found in the Pacific and Atlantic is absent in the Indian Ocean as there is no upwelling along the equator.

The SST decreases northwards during January- February (from about 28°C at 10° N to about 24°C in the northern most parts) in both the Arabian Sea and the Bay of Bengal. SST raises gradually as the Sun moves poleward over the northern hemisphere with the Bay of Bengal warming faster than the Arabian Sea. A zonal band of high SST (>30° C) is present in the south-eastern Arabian sea including the Lakshadweep Sea during April. SST exceeds 30° C in most of the area from equator to 15° N by May. With the onset of summer monsoon in June, SST starts falling and continues to fall until August due to strong winds and advection of upwelled water from the Somali and Arabian coasts. While the decrease of SST in the Bay of Bengal is only marginal (1° C), it is very prominent in the Arabian Sea and the equatorial part with a distinct spatial gradient toward west (SST< 25° C off the Somali coast). Once the southwest monsoon starts retreating in September, both the basins warm up until October. SST is more than 29° C in Bay of Bengal and is about 28° C in the Arabian Sea in October. Thereafter with the onset of northeast/winter monsoon in November, SST decreases in both the basins again till January. In the Head of the Bay of Bengal significant reduction in SST is noticed from October to December (29°-26°C).

The SST is found to vary from more than 28.5° C in the southern and eastern Arabian Sea to less than 18° C off the coast of Somalia. A remarkable feature of the Arabian Sea is that SST in July is 2° C lower than that in April, and in July, SST is almost as low as that in January. This typical behavior for northern latitudes in summer is caused by extreme ocean-atmosphere interactions during the southwest monsoon.

5.5.4. Seasonal Variation of SST at the Surface:

Northern winter (January-March): During the boreal winter, cells of warm water with temperatures around 29°C prevail off the south west coasts of India including the Lakshadweep and Male Islands, in the central Indian Ocean and off Sumatra coast in the east (Fig.5.21). The northward decrease of temperature in the Arabian Sea is more significant (around 4°C) than that in the Bay of Bengal. The region between the equator and 15°S has a temperature of about 28°C. South of 15°S, the temperature decreases southward and the isotherms are largely zonal with increasing meridional gradients towards higher latitudes. The orientation of these isotherms off the east coast of South Africa represents poleward moving western boundary current, the Agulhas Current, of the south Indian Ocean. The development of east Madagascar Current is also evident. In the eastern Indian Ocean surface waters warmer than 29°C are present in the Indonesian through-flow region. The presence of warm, poleward flowing eastern boundary current, the Leeuwin Current (shown in Fig.5.22), is evident from the orientation of isotherms off the coast of Western Australia.

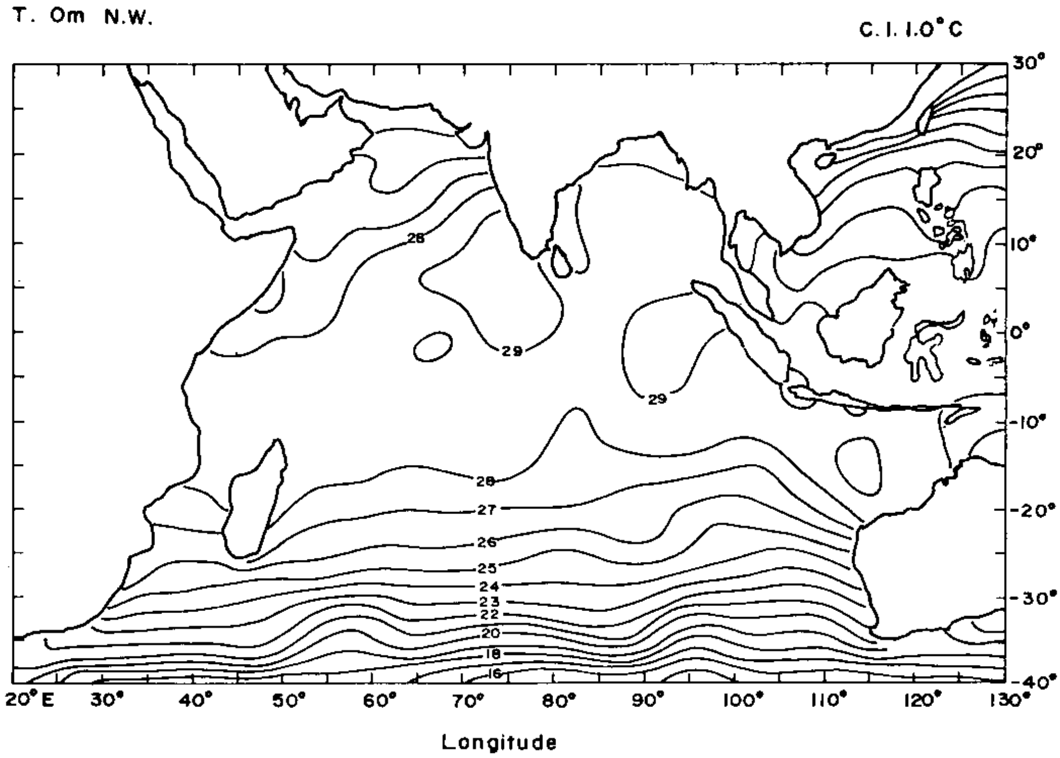


Fig.5.21. Temperature distribution at surface during Northern winter (January-March) in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram 2005)

Northern spring (April-June): During the boreal spring, the surface temperature increases to 29°C in the Arabian Sea, Bay of Bengal and the equatorial Indian Ocean up to the coast of Sumatra (Fig.5.23). Relatively cold waters are noticed off the coasts of northern Somalia and Arabia. In the south Indian Ocean, the isotherms are largely zonal, beginning from the east to 65°E and the east Madagascar, the isotherms exhibit divergence with the branching of the South Equatorial Current (SEC) towards the east African Coast and towards south-west Indian Ocean to feed the Agulhas Current (Fig.5.22). The orientation of the isotherms off the southern Africa shows the presence of Agulhas Current and its retroflection towards east to join the Antarctic Circumpolar Current (ACC).

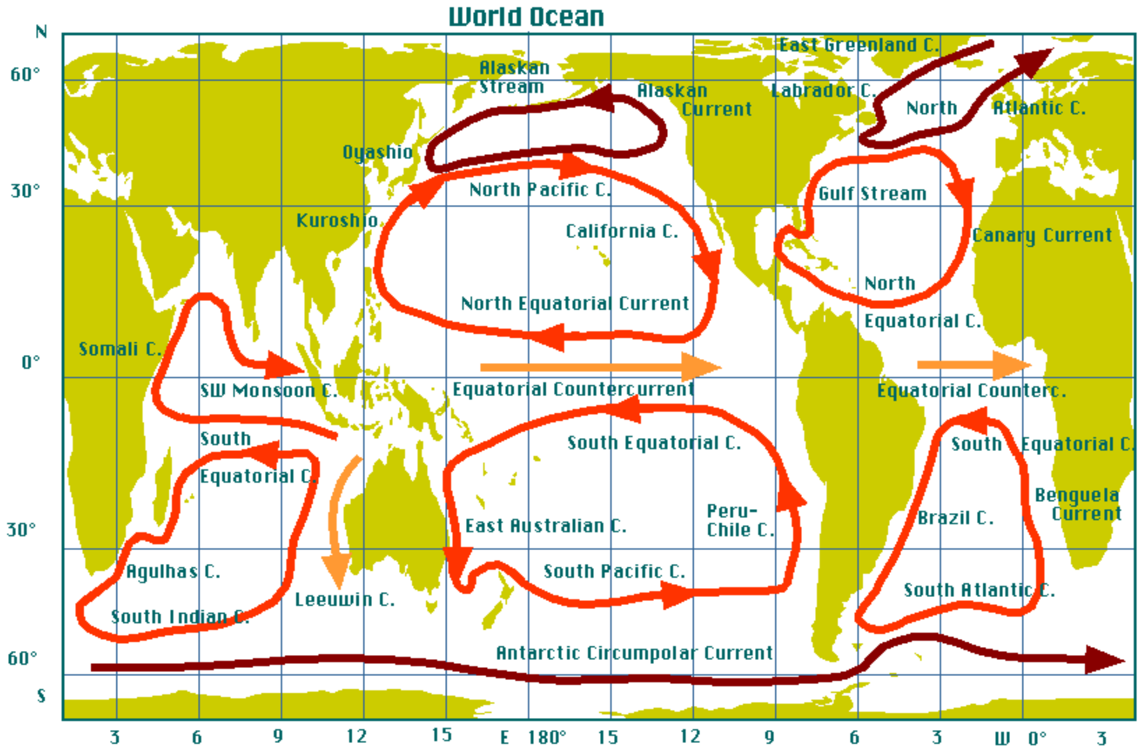


Fig.5. 22. A schematic of Surface currents of the world ocean during south west monsoon season.

A schematic of surface Currents in the Indian Ocean during south west monsoon are shown in Fig.5.22 for September/October (Southwest Monsoon season). The subtropical gyres are shown in red color, the subpolar gyres in North Atlantic, North Pacific and the Antarctic Circumpolar Current (equivalent current in the southern hemisphere) is shown in dark brown color. The Equatorial Counter Current and the Leeuwin Current are shown in orange color. Note that the figure does not indicate the strength of the current and the western boundary currents are much stronger than all other currents. Also please note the absence of Equatorial Countercurrent in the southern tropical Indian Ocean.

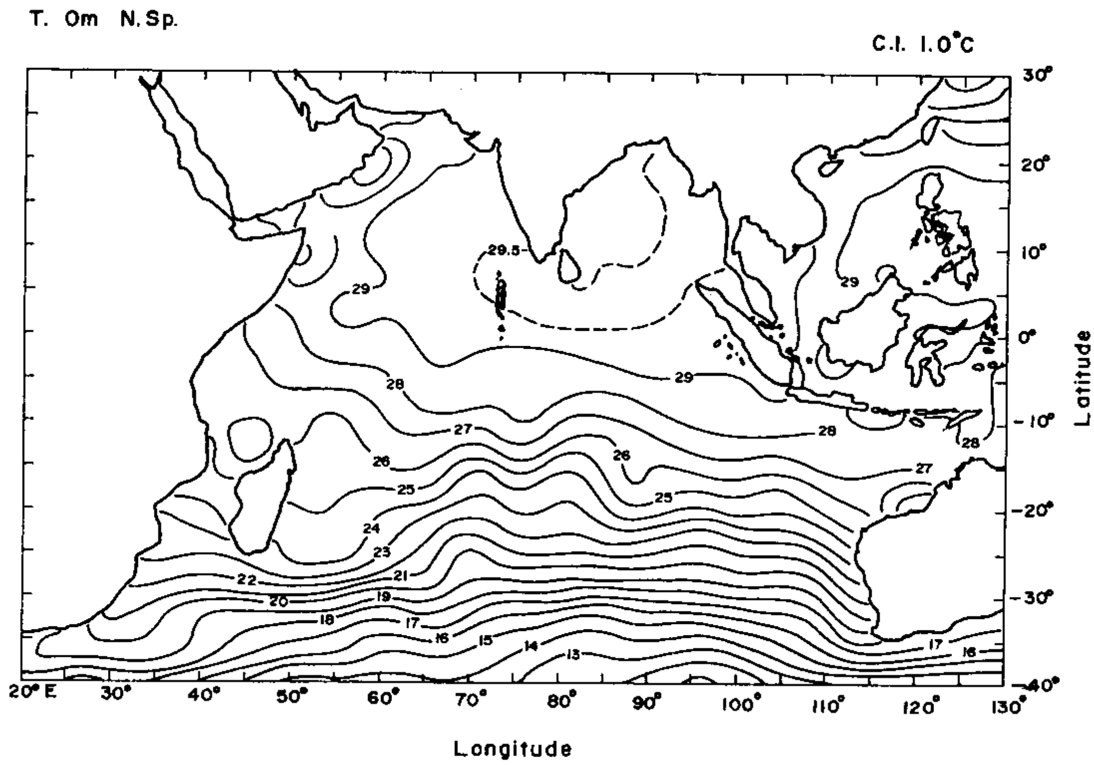


Fig.5.23. Temperature distribution at surface during Northern Spring (April-June) in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram, 2005)

Northern summer (July-September): During the boreal summer, the warmer (>28^{o}C) surface waters occupy the entire Bay of Bengal, the equatorial Indian Ocean up to 10^{o}S and southern Arabian Sea at 55^{o}E (Fig.5.24). Lateral temperature gradients are present to the west of 55^{o}E with the decrease of surface temperature towards the coasts of Somalia and Arabia under the influence of southwest monsoon winds and the associated intense coastal upwelling. Advection of colder surface waters towards the Central Arabian Sea (as far east as 65^{o}E) lowers the surface temperature appreciably there. In the south Indian Ocean, isotherms are zonal with a rapid decrease of temperature towards south. Northwards bifurcation of isotherms (>26^{o}C) at the northern Madagascar region, concordant with the equatorward branch of the SEC off the east African coast, is a noteworthy feature of this season. The Agulhas Current moving westward after rounding the peninsular region of South Africa is clearly evident.

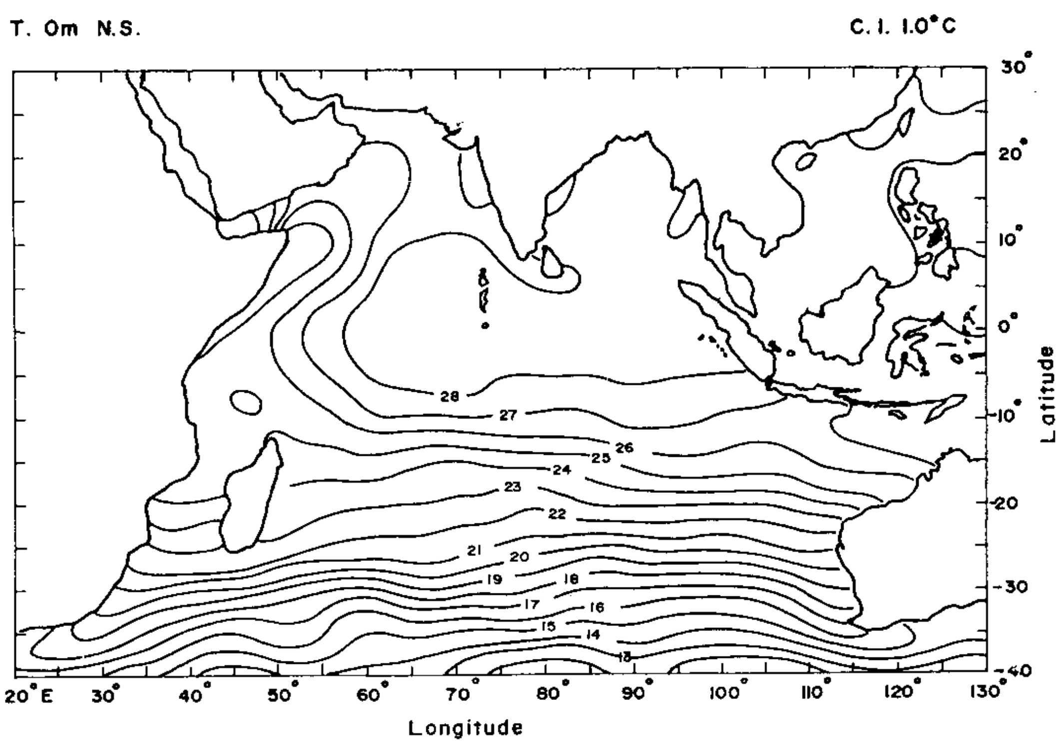


Fig.5.24. Temperature distribution at surface during Northern summer (July-September) in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram, 2005)

Northern fall (October-December): During the boreal fall season, the warmer (>28°C) surface waters are persistent in the equatorial region from the coast of Sumatra to 55°E and between 10°S and the southern Bay of Bengal (Fig.5.25). Surface temperature decreases towards northern Bay of Bengal, northern Arabian Sea, off the coasts of Somalia and Arabia and towards the southern Indian Ocean. South of 15°S, the isotherms align vividly in a NE-SW direction off the coasts of Madagascar and South Africa showing the presence of East Madagascar Current and the Agulhas Current.

5.5.5. Seasonal variation of temperature at 200 m level:

Northern winter (January-March): During the boreal winter, the temperature distribution at 200m depth is characterized by a narrow band of cold water (3.5-13.0°C) along 5-7°S and a broad band of warm water (19-20°C) along 20°S. This is followed by another band of cold water (11.5°C) at around 37°S (Fig.5.26). The strong meridional gradients occur both on the northern and southern side of this warm water band in the south Indian Ocean. The Agulhas current is well depicted with its retroflection in the offshore. North of 5°S, the temperature increases to 18.5°C gradually towards north in the northern Arabian Sea and to 15°C in the Bay of Bengal. The pattern of isotherms in the north Indian Ocean shows an equatorward flowing Somali Current and a gyre with clockwise circulation in the Bay of Bengal. The narrow cold water band at 5-7°S is due to the Ekman suction under the negative wind stress curl (in the S.H negative surface wind stress curl leads to divergence as against the surface convergence in the N.H). The warm water band on the south coincides with the southern limit of the SEC prevailing between 7°-18°S and the northern limit of the west wind drift current flowing eastward south of 20°S. These two current

systems together develop a broad zone of convergence centered at 20°S across the coast of Madagascar towards northern Australia. The warm temperature in the northern Arabian Sea is due to the influence of penetration of Persian Gulf water mass into the Arabian Sea in this depth domain.

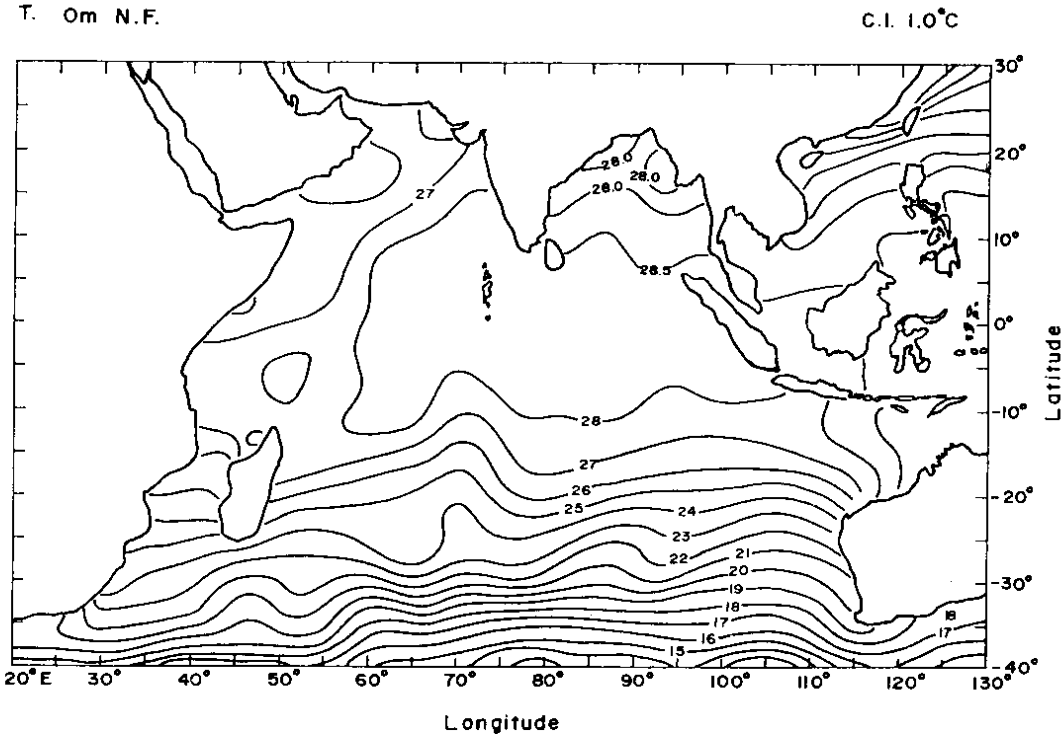


Fig.5.25. Temperature distribution at surface during Northern Fall (Oct-Dec) in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram, 2005)

Northern Spring(April-June): During the boreal spring, the temperature distribution at this depth does not change much from that during boreal winter, except for the orientation of the isotherms in the southern Indian Ocean (Fig.5.27). Off the coast of South Africa, meso-scale eddy- like structure is noticed, probably due to the retroflection of the Agulhas Current. There is evidence of the penetration of Persian Gulf water mass, at this depth towards the central Arabian Sea. The temperature in the Arabian Sea is higher than that in the Bay of Bengal due to penetration and spread of warm, saline PGW in the Arabian Sea.

Northern Summer (July-September): During the boreal summer, the penetration of the Persian Gulf water mass occurs chiefly towards central Arabian sea (12°N, 65°E) and from there towards the western Arabian Sea in a southwestward direction up to the equator (Fig.5.28). The variation of temperature in the Bay of Bengal, however, is smaller relatively. The other features remain the same as during boreal spring. Waters of nearly uniform temperature (13-13.5°C) dominate the southern Arabian Sea, the Bay of Bengal and the equatorial Indian Ocean up to as far as south as 5°S. A meso-scale eddy moves off shore from the Agulhas Current region.

Northern Fall (October-December): During the boreal fall season, the pattern of temperature variation is similar to that of the other seasons (Fig.5.29). This suggests that the temperature and

the associated flow patterns in the tropical Indian Ocean at the depth of 200m do not exhibit much seasonal variability.

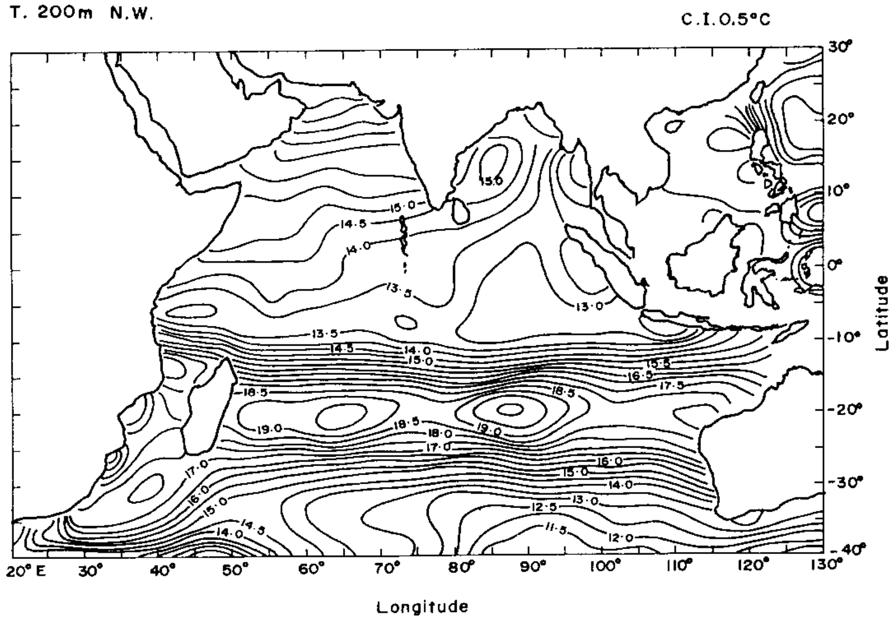


Fig.5.26. Temperature distribution at 200 m during Northern winter (January-March) in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram, 2005)

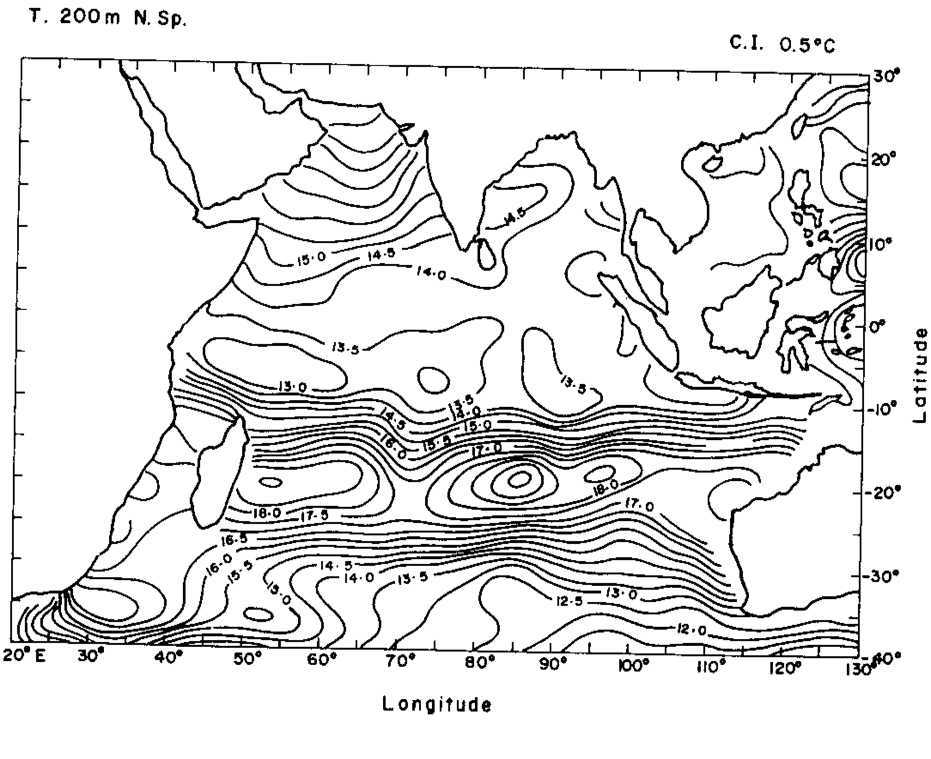


Fig.5.27. Temperature distribution at 200 m during Northern Spring (April-June) in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram 2005)

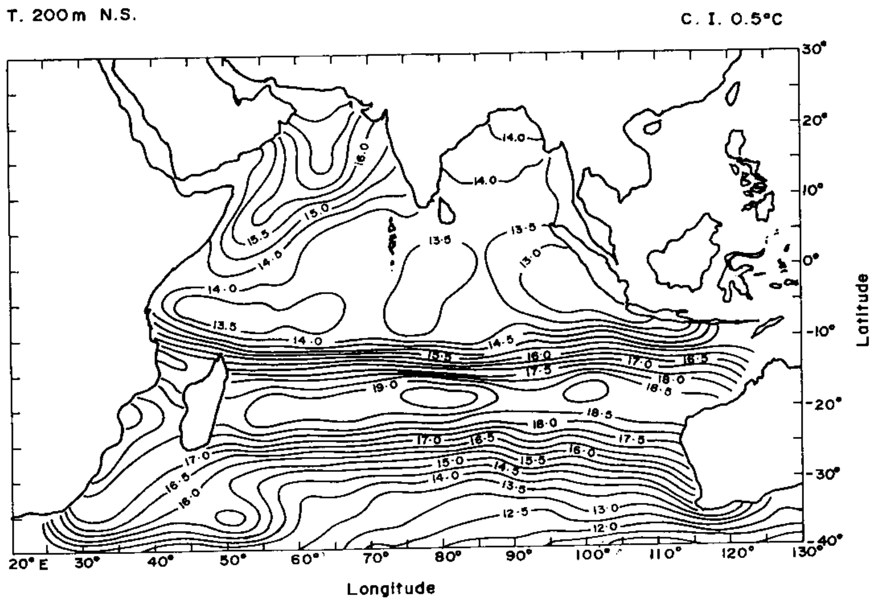


Fig.5.28. Temperature distribution at 200 m during Northern Summer (July-September) in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram, 2005)

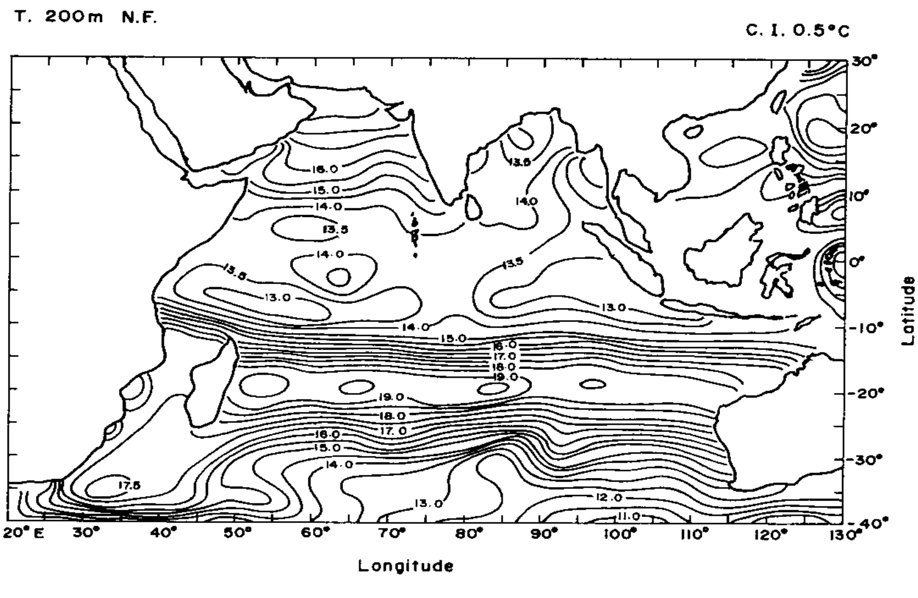


Fig.5. 29. Temperature distribution at 200 m during Northern Fall (October-December) in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram, 2005)

Annual Mean surface temperature:

On the annual basis, waters warmer than 29^{o}C occupy the entire Bay of Bengal, the Andaman Sea, off the southern tip of India and the east coast of Sri Lanka (Fig.5.30). The isotherms in the Arabian Sea exhibit SW-NE orientation with cold surface waters in the western Arabian Sea. The axis of the tongue of warm water lies along a line from the Bay of Bengal to the east African Coast (10^{o}S, 42^{o}E). South of the equator and east of 65^{o}E, the isotherms are largely zonal. The isotherm pattern diverges off Madagascar in association with the branching of the SEC such that one branch goes towards the east coast of Africa and the other one towards southern coast of Madagascar. This branching of the SEC is evident from the drifting buoys trajectories. The northern branch of the SEC feeds the east African Coastal Current and the southern branch feeds the Agulhas Current. Similarly, the isotherms off the coast of Western Australia bend poleward following the Leeuwin Current. The meridional gradients of temperature are rather large in the south Indian Ocean than north of the equator.

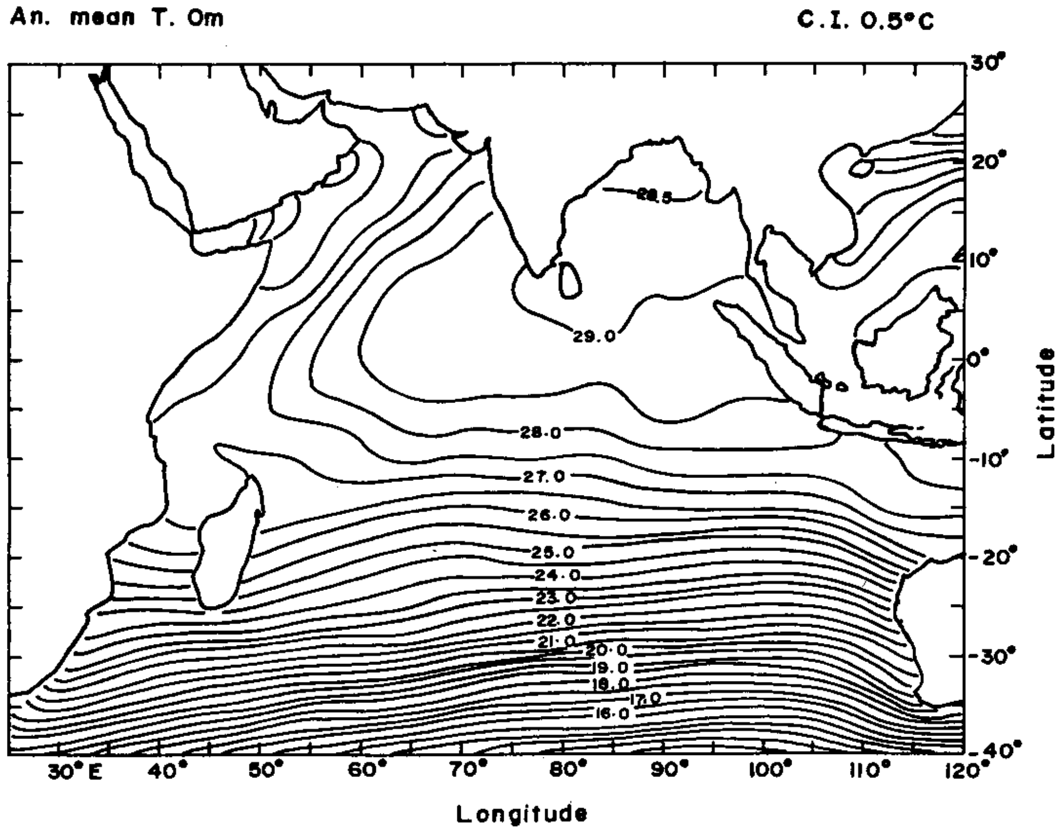


Fig.5. 30. Annual mean Temperature distribution at surface in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram 2005)

5.6. Indian Ocean Warm Pool (IOWP):

The minimum SST required for active convection is 28°C. A large region of tropical ocean with SST >28° C, called Warm Pool, lies in the western Pacific and the eastern and central Indian Ocean The area of Indian Ocean Warm Pool (IOWP) change three fold during a year from

a minimum of 8×10^6 km^2 during September to a maximum of 24×10^6 km^2 during April (Fig.5.31). The warming phase of IOWP begins in February, when the pool begins to spread on both sides of the equator. The northern boundary of the pool merges extends into the Bay of Bengal by April and into the Arabian Sea by May. The summer monsoonal cooling begins during June in the western equatorial Indian Ocean and during July in the Arabian Sea, and continues through September. In the Bay of Bengal the SST does not fall below 28° C during summer. The Warm Pool in the Bay of Bengal recedes during November – December. A study of SST data of 1992 revealed the persistence of Warm Pool (SST =28° C) in the Bay of Bengal from March through October.

The maximum depth of 28° C isotherm in the Indian Ocean is 100 m during May-June. The maximum volume of IOWP is 9×10^6 km^3 which is three times its minimum value. An important feature of IOWP is the large seasonal variation in the surface area covered by the pool. The dramatic changes that occur to the Warm Pool during April and May prior to the onset of the summer monsoon rains over India are particularly noteworthy. The factors that control the variability of IOWP include the fluxes of heat and momentum and ocean currents, whose spatial and temporal variation is complex and difficult to observe. An obvious conclusion is that the IOWP is a region of net annual heat gain by the ocean.

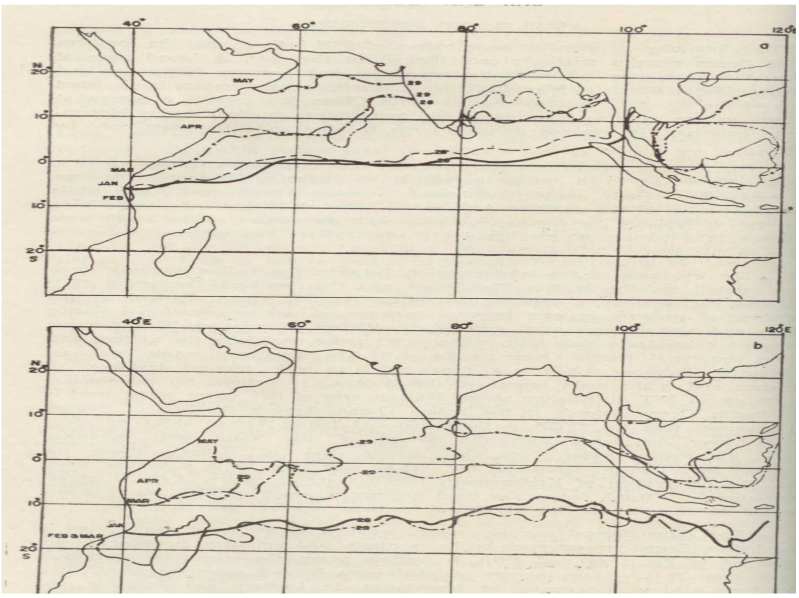


Fig.5. 31. The climatological mean position of warm water pool in the Indian Ocean for the period January to May. (a) the mean position of the northern boundary (b) the mean position of the southern boundary.(By courtesy of C.K.Gopinathan and D.Panakala Rao, 1985)

5.6.1. Arabian Sea Mini Warm Pool:

During the pre-summer monsoon (February-May) the near surface waters in the Arabian Sea progressively warm up and a mini warm pool with a core of temperature grater than 30° C is manifested in the southeastern region. This high SST (>30°C) develops off the Southwest coast of India in the Lakshadweep Sea in March well before the thermal equator moves into the area, and continues to retain its identity until the onset of monsoon. During most of the years the monsoon onset vortices have occurred only over the mini warm pool, suggesting that the SST high helps in

producing conditions favourable for the genesis of the monsoon onset vortex. The mechanism of formation of the SST high can be explained as below.

During winter the equatorward East India Coastal Current (EICC) forced by the downwelling coastal Kelvin waves in the Bay of Bengal, turns around Sri Lanka and on reaching the west coast of the Indian subcontinent flows as a poleward West India coastal Current (WICC) and carry low salinity waters from the Bay of Bengal to the Lakshadweep Sea. As the Kelvin waves propagate poleward along the west coast of India, they radiate downwelling Rossby waves that produce a high in the sea level off southwest India. The downwelling and the low salinity surface layer provide a breeding ground for the formation of a SST high in January and by March with increase in solar insolation and stable surface layer; the SST high is evident in the Lakshadweep Sea. Toward the end of April, the thermal equator runs over this high, further enhancing its temperature. By the end of May the SST high becomes the core of the warm pool over the Arabian Sea and there by becomes the warmest spot in that region.

The existence of a mini warm pool with temperature more than 30.25°C along 9° N between 68° and 75.5° E during May 2000 was also found. Existence of a clockwise gyre during May and the possibility of recirculation of low salinity waters in the study region are also inferred.

5.7. Horizontal distribution of temperature in the Bay of Bengal:

Sea surface temperature distribution in January showed a pattern, which was uniform in the zonal direction, but varied meridionally (Fig.5.32). SST was the coldest in the north with the minimum of 22^{o}C and increased steadily towards south where the maximum was 31^{o}C. In the western Bay average SST was seen about 26^{o}C, while in the eastern Bay, it was about 28^{o}c. In April, the SST showed a rapid heating of 3 to 6^{o}C in the northern Bay compared to winter. The spatial variation is only marginal from 29^{o}C to 31^{o}C in the entire Bay, except in small pockets along the eastern boundary where it was 32^{o}C and in the northern most coastal region where it was 28^{o}C. By July, as the summer monsoon sets in the Bay, the SST was more or less uniform in the entire Bay with a value of 29^{o}C (Fig.5.33). However, SST of 30^{o}C was noticed in the southwestern and southern region, while 31^{o}C was seen in the southeastern region. Pockets of colder waters with SST 28^{o}C were seen in the central and eastern Bay. In August, the SST pattern remains more or less similar to July, except the occurrence of colder waters around Sri Lanka and northeastern Bay intruding into the central Bay. In October, in the northern Bay SST was warmest of all the months with a value of 31^{o}C. Entire Bay was very warm with SST of 30^{o}C, except in a zonal patch in the south and a small pocket near the western boundary, which was 1^{o}C colder. The distribution pattern of SST during October was just reverse of that during January. See Table 5.1 for the basin-averaged SST for each of the above-discussed month along with the minimum and maximum values.

However, the climatological distribution of temperature shows much less variability except during winter (Figure 5.34).

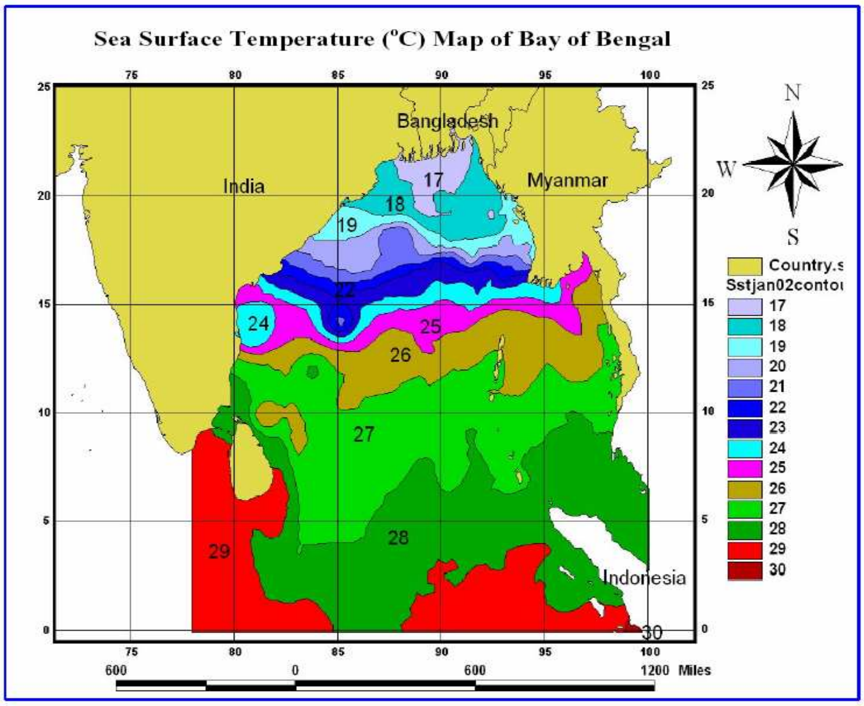


Fig.5.32. Monthly mean SST (^{O} C) distribution in Bay of Bengal in January (By courtesy of BalaKrishna Patidar, 2006)

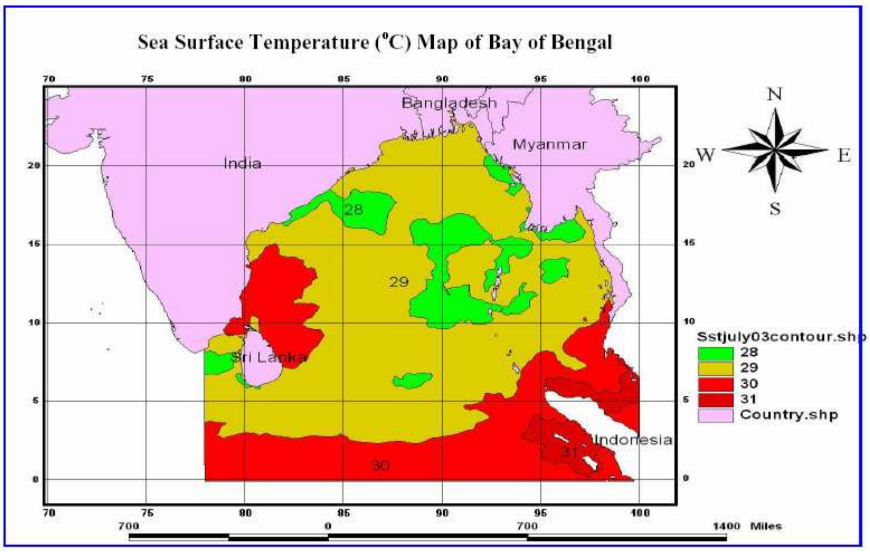


Fig.5.33. Monthly mean SST (^{O} C) distribution in Bay of Bengal in July (By courtesy of BalaKrishna Patidar, 2006)

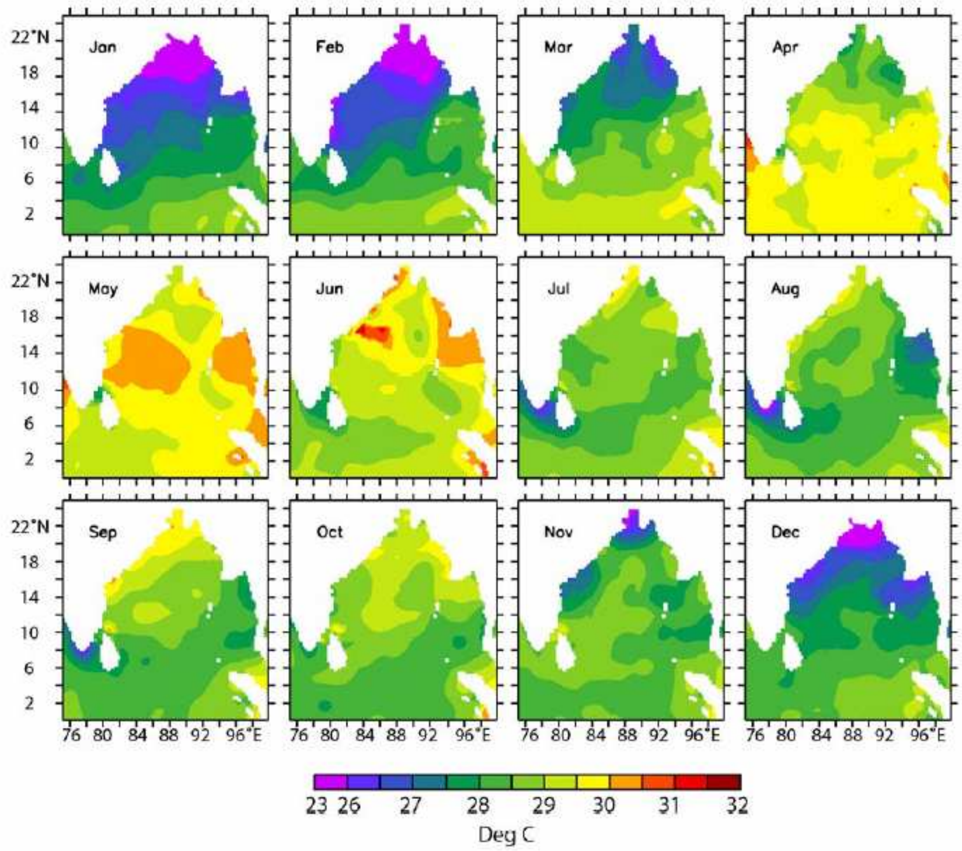


Fig.5.34. Sea surface temperature (^{O} C) in Bay of Bengal obtained from the WOA 01 climatology (Conkright, et al., 2002)

Month of 2003	Sea Surface Temperature (^oC) in 2003		
	Minimum	Maximum	Average
January	21	31	28
April	27	32	30
July	27	31	29
August	26	31	28
October	28	31	29

Table 5.1. Basin averaged SST (^{O} C) with its maximum and minimum value during different months in the year 2003

5.8. Distribution of temperature in the coastal regions of Bay of Bengal:

During the winter monsoon of 1983 surface temperature (Fig.5.35a) in the coastal regions of Bay of Bengal increased from 25^{o}C at 18^{o}N to 28.5^{o}C towards the south off Chennai and to 26^{o}C at about 20^{o}N. At 50m (Fig.5.35b) the temperature varied from 26^{o}C at 16^{o}N to 28^{o}C towards both the south and north. A few pockets of higher temperatures compared with the surface, were observed at this depth. Balaramamurty (1958) and Rao and sastry (1981) attributed this to surface cooling. At 100m during the winter monsoon a broad cold-water cell (<17^{o}C) appeared between 13^{o}N and 16^{o}N (Suryanarayana et al. 1993). At 200m also (Fig.5.35c), below the thermocline, the lowest temperature (13^{o}C) was recorded around 16^{o}N, increasing towards the south and north. During the winter monsoon the sea surface temperature in the southwestern region was high compared to that in the other regions along the east coast. The thermocline first dips to about 100 m and then becomes shallower towards offshore. Off Visakhapatnam, the thermocline is deeper and becomes uniformly shallower from 100-200m. Further northwards, off Mahanadi, also the thermocline shows in a similar fashion. At the head of the Bay, off the coasts of Myanmar and Bangladesh, the surface temperature is 26^{o}C as is that off Chilaka Lake. Here, as the shelf is wide the isotherms dip towards deeper regions. Off the Irrawady towards the west and south the surface temperatures range between 27.5^{o}C and 28^{o}C. A sharp thermocline is noticed in the west-east section where the maximum depth is about 80m. During Jan-Feb of 1983, Janekarn &Hylleberg (1989) found low temperatures between 25.6^{o}C and 27.2^{o}C in the coastal waters off the west coast of Thailand. They found that bottom temperatures varied greatly, in some places between 25.0^{o}C and 26.7^{o}C and in other places between 27.5^{o}C and 28.3^{o}C. They also observed thermocline depths varying between 40m and 45m in the coastal and offshore areas (from 9^o 5'N to 6^o 30'N) during Jan-April. These surface temperatures are about 2^o C lower than those

observed off the Chennai coast on the west side of the Bay. Similarly these thermocline depths are also much lower than those observed between 18°N and 21° N along west coast of Bay of Bengal. In the entire east coast of India at various places inversions are found during winter. These inversions are considered to be highly important in terms of over turning of surface water mass and instability. The thickness of the mixed layer extends up to 60m outside the shelf area in the northern Andaman Sea.

During the summer monsoon (Fig.5.36a) surface temperatures vary between 28°C and 30°C. In the northern coastal belt temperature difference between winter to summer is 4°C (26°C to 30°C), whereas off Chennai the temperature remains at 28°C. At 50 and 200m surfaces (Fig.5.36 b& c) due to the fluctuations in the thermocline, pockets of high and low temperatures are found. Off Chennai a shallow thermocline of 50m is observed. Off the central east coast of India the thermocline shows wavy pattern and a warm water cell (23°C) is also found at about 100m. Off the northern coast though surface temperatures increase from winter to summer, the thermocline variation is very little. Surface temperatures of Andaman Sea are around 29°C near Pucket Island (Sandstrom et al, 1987) during April-May. Similar temperatures were noticed in the southern east coast of India also. The low temperature cell (<25°C) observed at 50m depth (Fig.5.36b) during summer represents a cold core eddy according to Babu et al (1991). They found that the eddy has temperature drop of 4 to 5°C at its centre and extends between 50 and 300db surfaces with a diameter of about 200 km. Eddies are generally generated by baroclinic instability at the interface of two opposing currents present along the shelf edge.

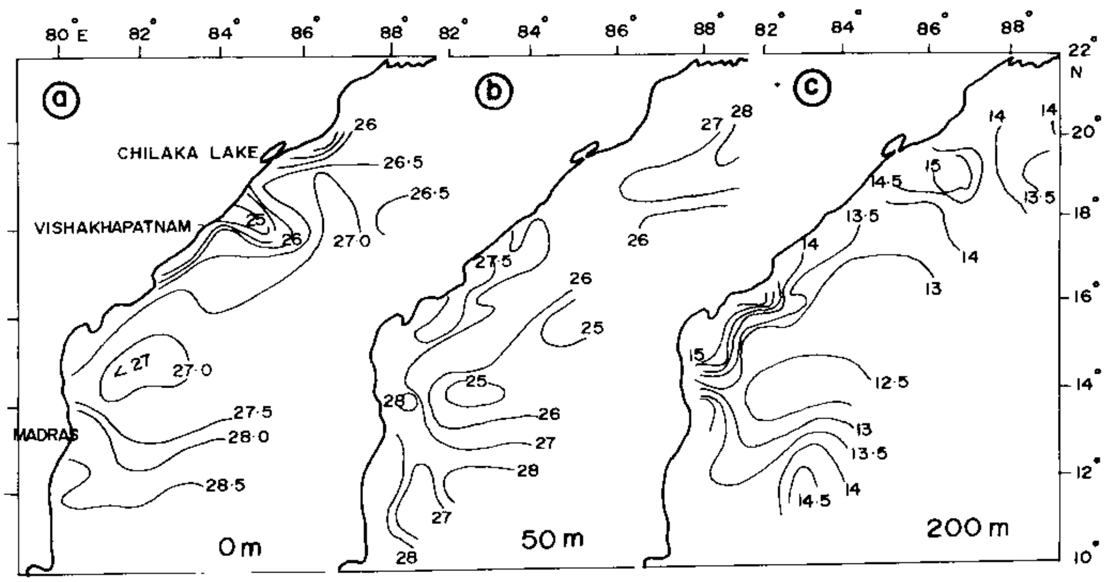


Fig.5. 35. Distribution of temperature during winter at various levels in coastal areas of Bay of Bengal. (By courtesy of Suryanarayana et al, 1992)

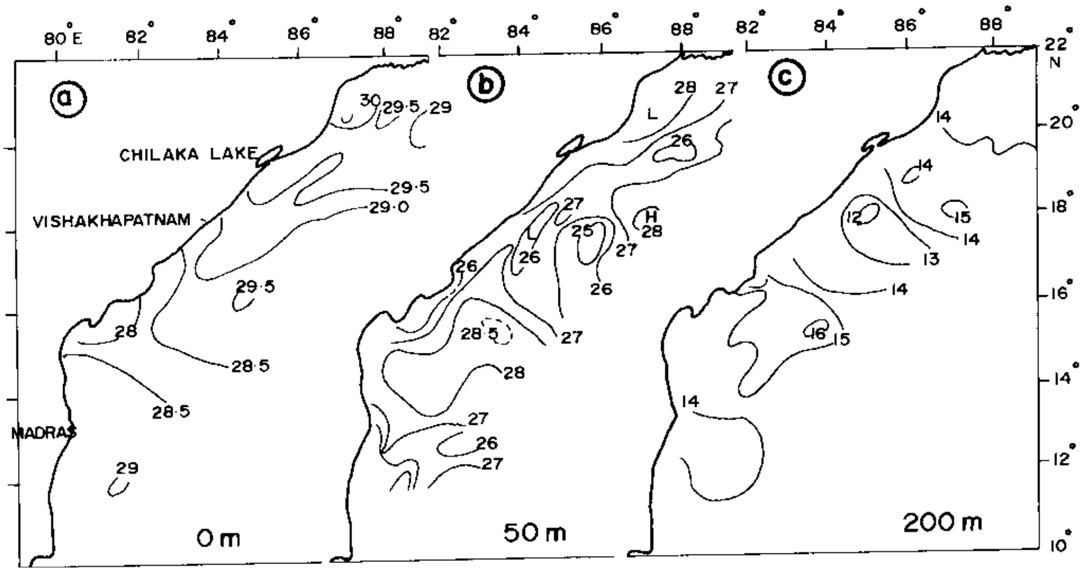


Fig.5. 36. Distribution of temperature during summer at various levels in coastal areas of Bay of Bengal. (By courtesy of Suryanarayana et al., 1992)

5.9. Distribution of Salinity:

The word *salinity* is defined as the total amount of dissolved material in grams present in one kilogram of sea water when bromine and iodine are replaced by equivalent amount of chlorine, when carbonates are oxidized and all the organic matter also is oxidized.

The total salt content presently dissolved in the ocean is about 5×10^{19}Kgs and every year nearly 2.73×10^{12} Kgs of salt is added to the sea.

Although the average salinity is put at 34.7‰, numerous salinity differences are encountered in horizontal as well as in vertical direction. The factors that influence salinity in the oceans are as below:

Increase	Decrease
1. Evaporation	Precipitation
2. Ice formation	Ice melting
3. Intrusion of high saline water(advection of high saline water)	Advection of low saline water.
4. Mixing with more saline deep water (turbulence & convection)	Mixing with less saline deep water
5. Dissolving of new salts due to deep sea oozes and sub marine volcanic eruptions.	In flow of fresh water due to rivers, glaciers and icebergs.

The average surface salinity distribution is shown in Fig.5.37 (in the lower graph in red). It has a minimum just north of the equator and maxima are found in the subtropics (30^o N & S) and later decreases to high latitudes. Observations make it clear that surface salinity variation depends on (E-P) factor. Here 'E' is evaporation and 'P' precipitation. If (E-P) is positive in an area, salinity will increase and if (E-P) is negative, it decreases salinity. The salinity maxima in subtropics may be because evaporation exceeds precipitation, while the equatorial minima are due to precipitation increases evaporation.

You may note here Evaporation, precipitation and river run-off are expressed as volume per unit time. Modern oceanography uses more and more a unit called "sverdrup" (Sv), defined as 1 million cubic meters per second: 1 Sv = 10^6 m^3 s^{-1}. The conversion factor for km^3 $year^{-1}$ to Sv is 1000 km^3 $year^{-1}$ = 0.0317 Sv. This gives an evaporation of 14.0 Sv, a precipitation of 13.1 Sv and a total river run-off of 0.9 Sv.

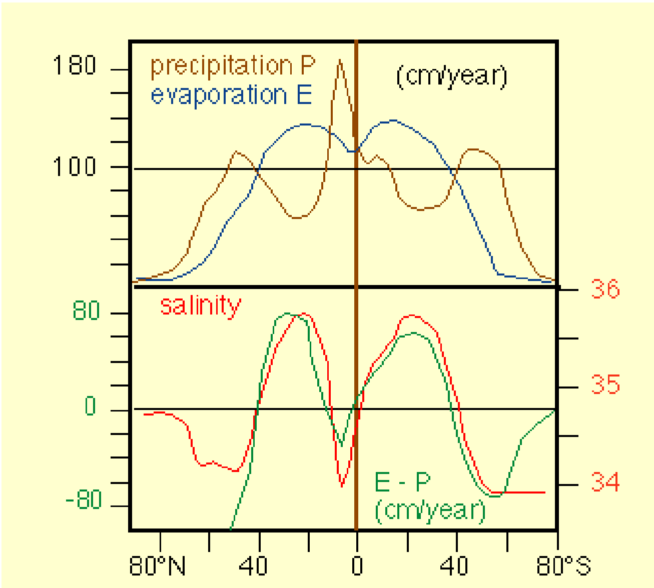


Fig.5.37. Annual mean distribution of evaporation *E*, precipitation *P*, the difference *E* - *P* (left scale), and sea surface salinity (right scale) in red (By courtesy of www.incois.gov.in/Tutor/IntroOc/notes/figures/fig4a5..html).

On a global scale, the balance is like this: Evaporation = $440 \cdot 10^3$ km^3 $year^{-1}$, Precipitation = $411 \cdot 10^3$ km^3 $year^{-1}$, River run-off = $29 \cdot 10^3$ km^3 $year^{-1}$. The distribution of sea surface salinity mirrors the distribution of E-P over large parts of the ocean (Figure 5.30). Deviations occur from river run-off.

The range of surface salinity values in the open ocean is from 33 to 37‰. Lower values occur locally near the coasts where large rivers empty and in the Polar Regions where ice melts. Higher values occur in the regions of high evaporation such as the eastern Mediterranean Sea (39‰) and the Red Sea (41‰) and the Persian Gulf (40‰). On an average the North Atlantic is the most saline at the surface (35.5‰), the South Atlantic and the South Pacific less (35.2‰) and the North Pacific the least saline (34.2‰) among the major oceans.

5.9.1. Horizontal distribution of sea surface salinity (SSS) in TIO:

During January to June, the SSS is more than 36‰ in the Arabian Sea, north of 10^O N and west of 70^O E, and decreases towards SE, where as in the Bay of Bengal, the SSS increases from about 30‰ in the head of the Bay towards South up to about 34 ‰ at 5° N. The SSS distribution in January clearly shows the spreading of low salinity waters (Salinity less than 34‰) into the southeastern Arabian Sea under the influence of northeast monsoon. During July to December, the SSS of more than 36 ‰ in the Arabian Sea (between 50° and 60° E) extends towards south up to equator, while in the Bay of Bengal the SSS increases significantly from less than 28‰ near the coast to about 34‰ at 10° N.

In the equatorial region, the SSS decreases from more than 35‰ in the western part to less than 34 ‰ in the east reflecting the influence of low salinity waters of the Indonesian through flow. South of 10° S, the SSS distribution is mainly zonal and increases towards south reaching maximum (of more than 35.5‰) around 30° S reflecting the P -E minimum in this region.

5.9.2. Salinity of Bay of Bengal and Arabian Sea:

The Bay of Bengal (BOB) is markedly different from the Arabian Sea in two respects. Firstly, over the Arabian Sea, evaporation exceeds precipitation and it receives higher salinity water from the Red Sea and the Persian Gulf, making the salinity of upper layer of Arabian Sea higher. In contrast, over the Bay of Bengal rainfall exceeds evaporation and also it receives large quantities of fresh water from the major rivers that drain into Bay of Bengal. This situation makes the upper layers of Bay of Bengal less saline for a large part of the year. Hence the salinity of surface waters of the Bay of Bengal is so low that the stratification in the upper layer is dominated by salinity gradients rather than the temperature gradients.

The huge quantity of river runoff coupled with the excess precipitation induces large changes in the salinity, which is predominantly felt along its peripheries (Figure 5.38). Low salinity with comparatively weak winds makes the BOB a highly stratified basin.

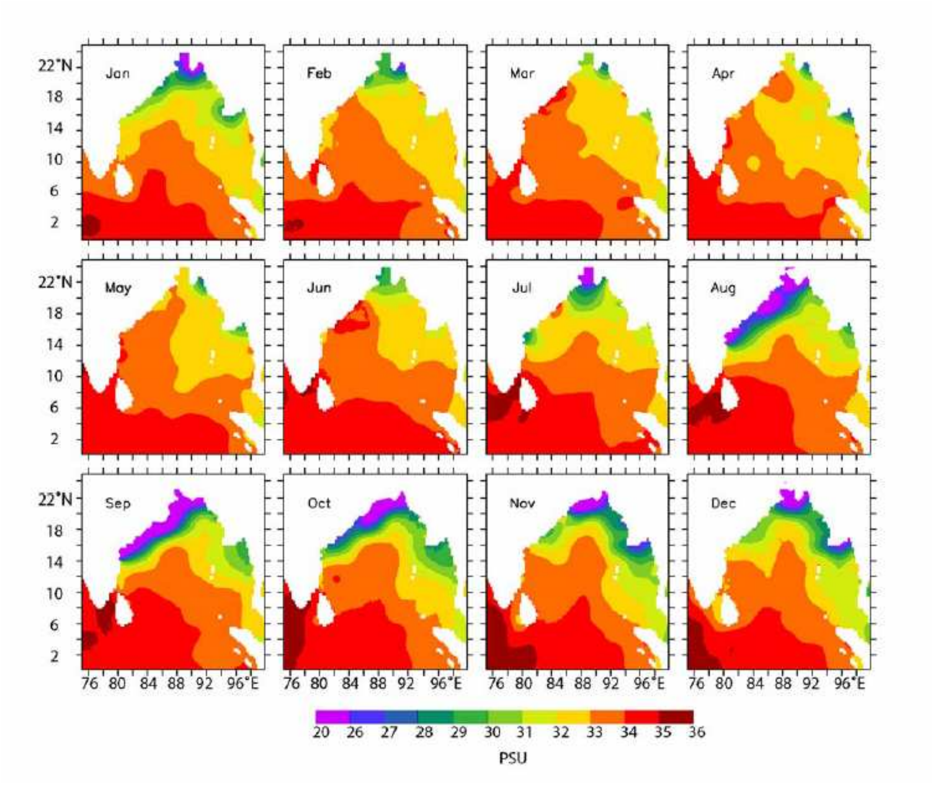


Fig. 5.38. Surface salinity in the Bay of Bengal (BOB) obtained from WOA01 climatology. Note the northern regions of the BOB is least saline through out the year than the southern BOB due to high river discharge (Conkright et al., 2002)

5.9.10. Seasonal variation of surface salinity in TIO:

NORTHERN WINTER (Jan-March):

During boreal winter, the sea surface salinity (SSS) in the Arabian Sea and Bay of Bengal differ very much (Fig.5.39). While higher salinity (> 36.4‰) dominates the northern Arabian Sea, lowest salinity (< 32.0‰) occurs close to the head of the Bay of Bengal. This clearly reflects the dominant control of evaporation over precipitation in the Arabian Sea and the major contribution of river run-off and precipitation over the evaporation in the Bay of Bengal. The higher SSS and the lower Sea Surface temperatures (SSTs) in the northern Arabian Sea contribute together to the increase of surface water density and lead to the formation of Arabian Sea High Salinity Water mass (ASHSW). This ASHSW sinks to subsurface depths (50-100 m) and spreads southward in the Arabian Sea and Equatorial region. This is further advected eastward by the seasonal Indian Monsoon Current into the central Bay of Bengal. While the isohalines are aligned in a NE-SW direction in the Arabian Sea, they exhibit south east orientation in the Bay of Bengal. This feature leads to the eastward decrease of salinity in the equatorial region. Across the gap between northern Australia and Java, low salinity waters of the Indonesian through-flow penetrate towards west and as a result a tongue of low salinity (< 34.2‰) appears in the southern tropical Indian Ocean between 8^{o}S and 15^{o}S. A zone of high salinity (>35.8-36.0 ‰) lies along 35^{o}S in the southwest Indian Ocean and shifts towards 30^{o}S in the east, off the coast of Western Australia. This salinity tongue results from the higher evaporation under the influence of strong westerly winds. This tongue of high salinity describes the southern subtropical anticyclonic gyre. These high salinity waters are associated with lower temperatures (11.5-14.5^{o}C) (Fig.5.21) and leads to the formation of Indian Ocean Central water mass which forms the major source for the ventilation of thermocline in the south Indian Ocean. The pattern of isohalines off the coast of South Africa shows that the Aghulas Current transports relatively saline (>35.4‰) waters of the Arabian Sea origin.

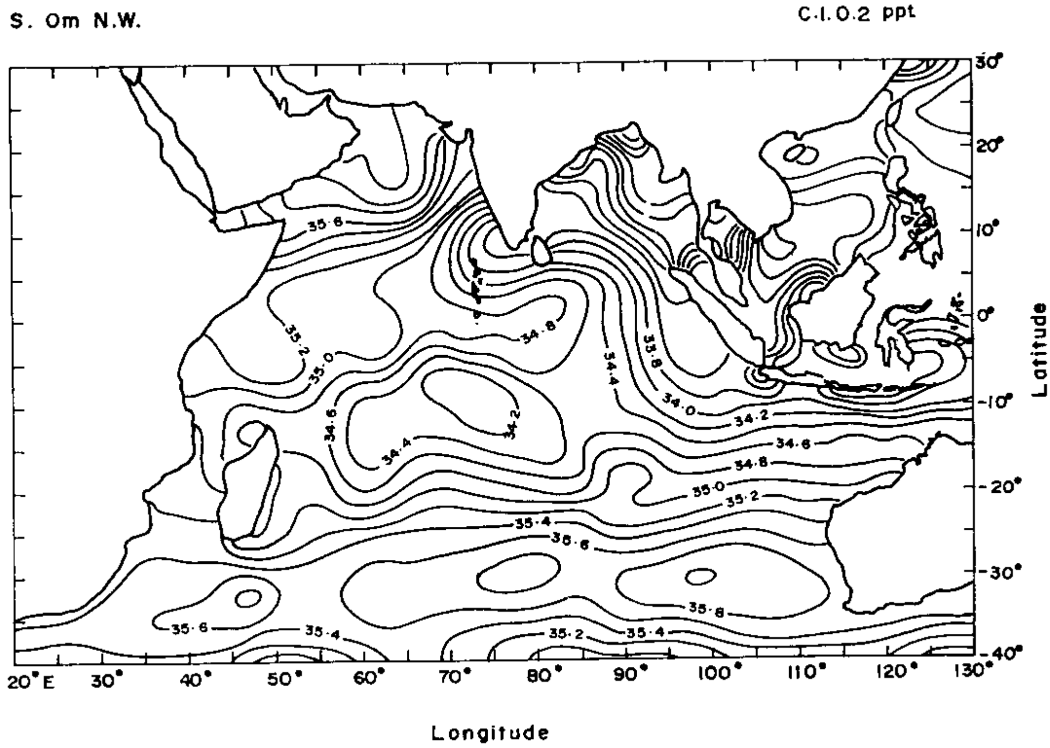


Fig.5.39. Salinity distribution at sea surface during Northern winter (Jan-Mar) in TIO. (By courtesy of Rao L.V.G and P.Shri Ram 2005)

NORTHERN SPRING(April-June):

During the boreal spring season also the Bay of Bengal exhibits the lowest salinity compared to the Arabian Sea, though the salinity of the south western Bay increases from that during boreal winter (Fig. 5.40). The low salinity cells are confined to the areas of river mouths. Strong gradients can be noticed in the northern Andaman Sea and northeastern Bay of Bengal. During this period, isohalines in the Arabian Sea and the Bay of Bengal orient towards SW and SE respectively, and the salinity in the equatorial region decreases eastward. In the region of South Equatorial Current (SEC), the isohalines are zonal and orient southwestward up to the Aghulas Current. The salinity of the zone of salinity maximum is relatively high off the coast of Western Australia and this zone is located at 30°S off the Western Australia than that at 37°S off the South African coast.

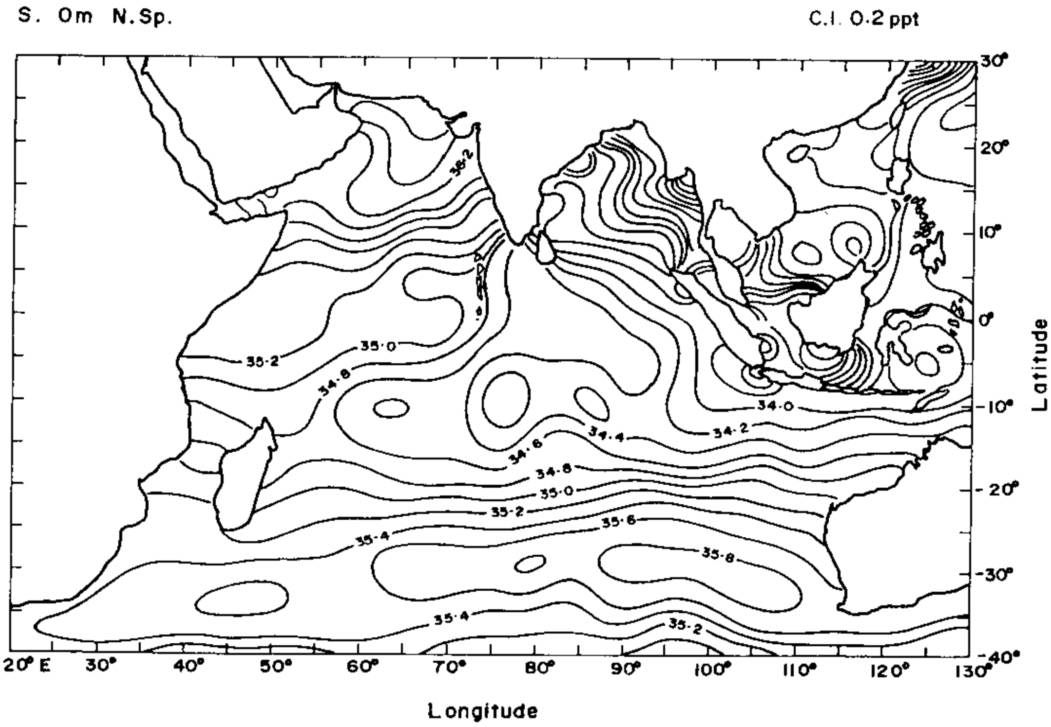


Fig.5. 40. Salinity distribution at sea surface during Northern spring (April-June) in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram 2005)

NORTHERN SUMMER(July-Sept):

During the boreal summer, the lowest salinity waters influenced by huge freshwater influx during the south west monsoon (from the rivers the Ganges, the Brahmaputra etc.) are seen in the western Bay of Bengal (Fig.5.41).

The southeastward orientation of the isohalines from the central Bay up to the Java coast is still evident. In the Arabian Sea, though salinity distribution is invariant, the pattern of isohalines suggests a flow of highly saline waters towards southwestward and southward directions. The difference in salinity of the equatorial Indian Ocean decreases from that during boreal winter and spring. The flow of low salinity waters through SEC is limited up to 55°E. The zone of subtropical maximum salinity is persistent with the higher salinities in the east.

NORTHERN FALL(Oct-Dec):

During the boreal fall, the region of lowest salinity in the Bay of Bengal shifts to the northeastern Bay and cells of low salinity occur at the river mouths (Fig.5.42). The isohalines in the southern Bay still persist towards southeast. In the Arabian Sea, the high salinity waters exhibit a tongue like distribution with its axis directed southward in the central Arabian Sea. The pattern of isohalines shows eastward flow of high salinity waters north of equator and westward flow of low salinity waters at around 10°S. South of 20°S the isohaline pattern is persistent from the earlier seasons.

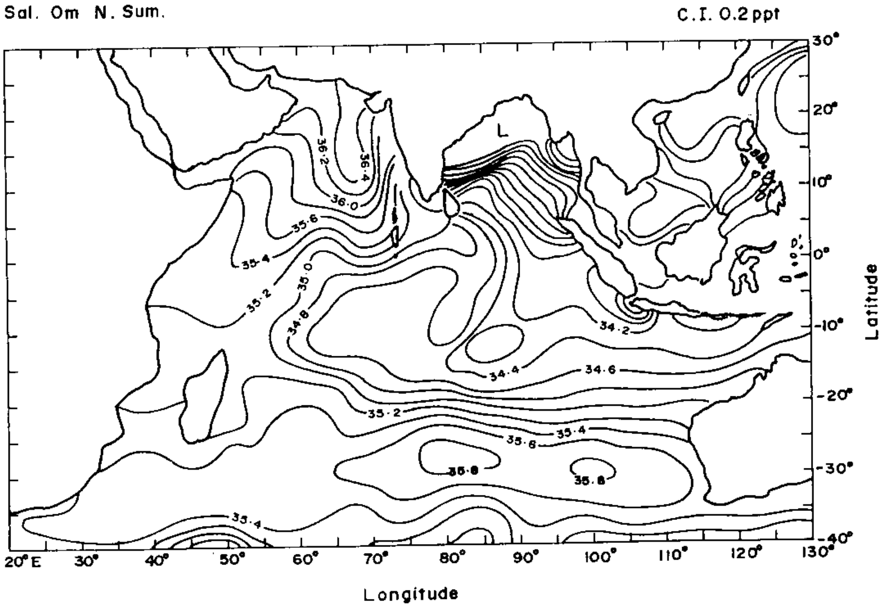


Fig.5. 41. Salinity distribution at sea surface during Northern Summer (July-Sept) in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram 2005)

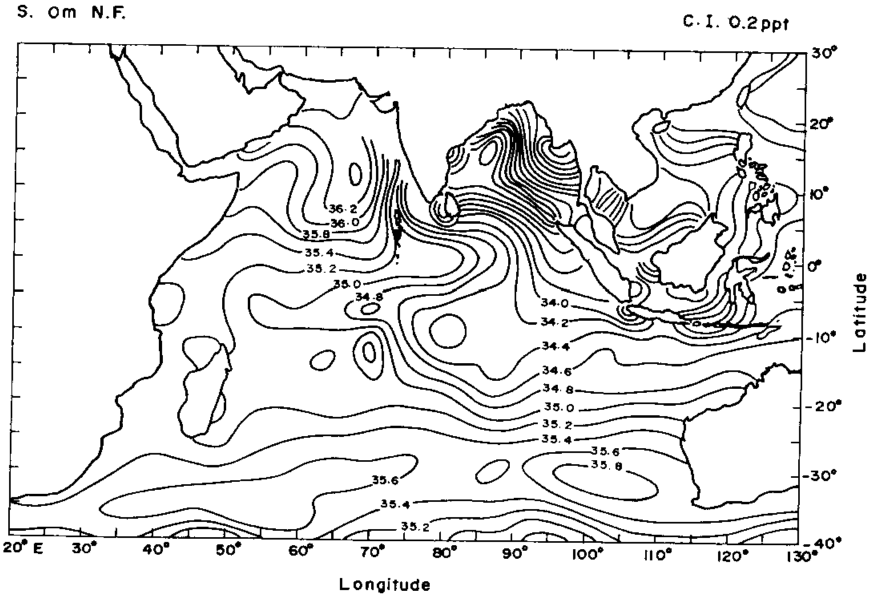


Fig.5. 42. Salinity distribution at sea surface during Northern Fall (Oct-Dec) in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram, 2005)

ANNUAL MEAN SURFACE SALINITY:

On the annual scale, the waters with highest salinity are present in the northern Arabian Sea and with lowest salinity in the western, northern and eastern part of the Bay of Bengal (Fig.5.43). The isohalines in the southern Arabian Sea and the southern Bay of Bengal exhibit SW-NE orientation in the former and NW-SE orientation in the latter (up to the coast of Java in the south East Indian Ocean). Low salinity waters from the south-east Indian Ocean appear to flow up to the south west coast of India. The isohaline pattern in the equatorial region indicates the eastward penetration of high salinity waters along 3°N and west ward penetration of low salinity waters around 10°S. Large east-west salinity differences are noticed in the 5°N-5°S latitudinal belt with higher salinity off the east coast of Africa. South of 10°S, salinity increases southward to a maximum in the subtropical high salinity zone. The isohalines are largely zonal. The maximum salinity of the subtropical high salinity zone also exhibits an eastward increase with the maximum occurred off the southern coast of Western Australia. The isohaline pattern further shows the presence of east Madagascar Coastal Current, the Aghulas Current and the Leeuwin Current, all transporting low salinity waters towards higher latitudes.

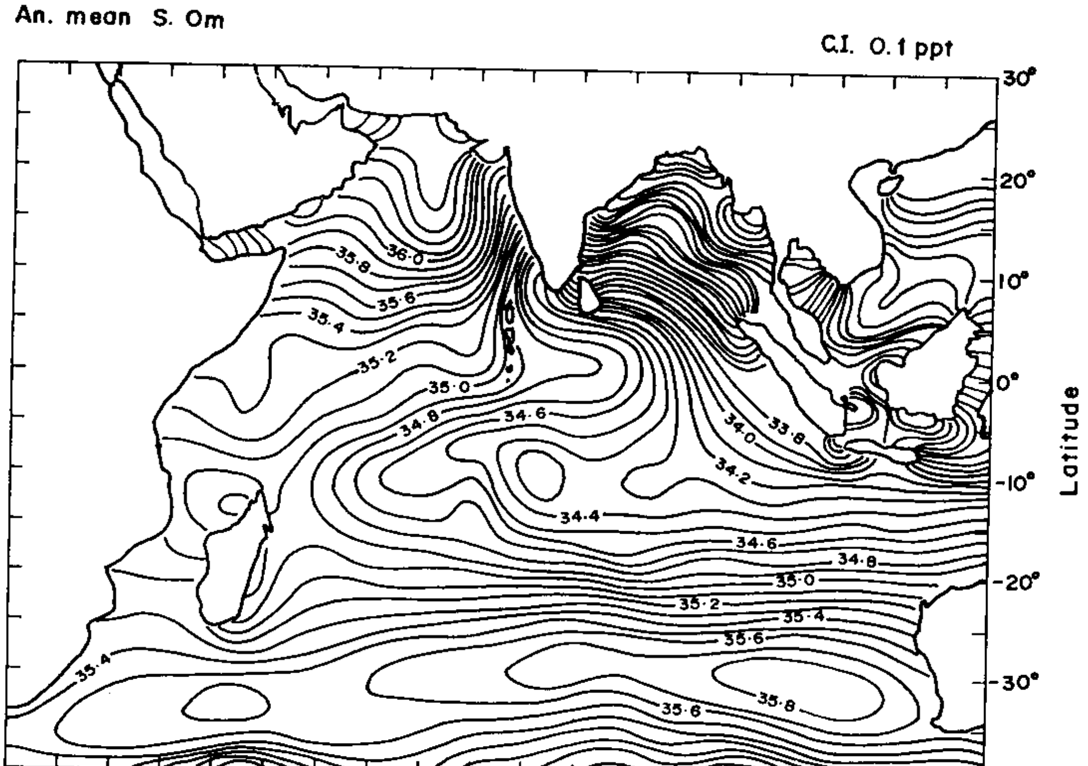


Fig.5. 43.Annual mean sea surface salinity distribution in TIO. (By courtesy of L.V.G.Rao and P.Shri Ram, 2005)

5.8.3. Vertical distribution of Salinity in the Arabian Sea and Bay of Bengal:

The vertical distribution of salinity is complex compared to that of temperature. This is because the density which determines the stability of the water body depends more on temperature than on salinity in the open ocean, except in the polar seas. Hence it is possible to have either high or low salinity in the upper warmer waters. In the equatorial and tropical regions of the oceans there is a uniform layer of salinity from the surface up to a few tens of meters depth due to mixing by wind and waves. Below this layer, salinity first increases up to 100 to 200 m and then decreases with depth reaching a minimum around 800 - 1200 m. The zone or layer where salinity changes rapidly with depth is called the halocline.

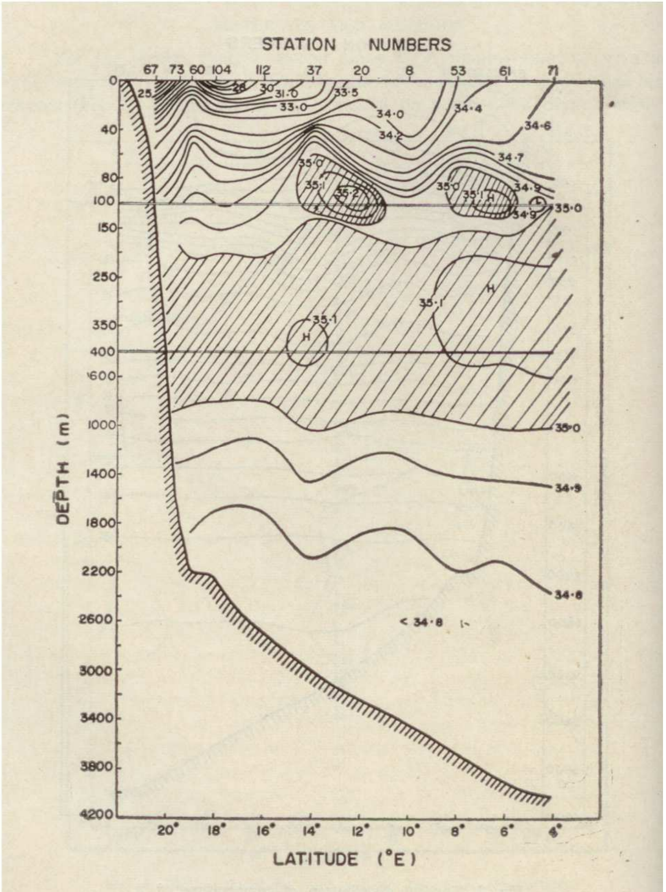


Fig.5. 44. Vertical distribution of salinity (‰) in the Bay of Bengal. (By courtesy of J.S..Sastry et al. 1985)

The surface salinity along 88°E in the northern Bay of Bengal (Fig.5.44) shows strong gradients in the upper 100m. The salinity at the surface increases from about 21.0‰ in the northern regions to 34.6 ‰ in the southern regions, the gradients being stronger in the northern Bay of Bengal. Two isolated high salinity pockets could be seen centered around a depth of 100m. These two isolated high salinity pockets are interpreted as the remnants of the Arabian Sea High salinity water mass. The salinity maxima in the depth ranges of 200-900 m have been explained as due to the penetration of the isohaline layer which forms when Persian Gulf and Red Sea water masses come in contact in the Arabian Sea.

In the northern Arabian Sea the surface salinity is high (>36.0‰) through out the year due to increased evaporation and low humidity. There is a subsurface salinity maximum in the

upper 200 m corresponding to Arabian Sea high salinity (ASHSW) water. Its value varies from >36.7 ‰ around 20 m depth in the northern Arabian Sea to about 35.3‰ around 100 m depth in the south. A salinity maximum associated with the core of Persian Gulf water is found with value varying from 36.8‰ at about 200 m in the Gulf of Oman to about 35.0 ‰ at around 400 m in the south. Another salinity maximum associated with the core of Red Sea water (RSW) is found with value varying from 35.6‰ in the north to about 35.1‰ near the equator. A salinity minimum is often found at a depth of 400 m in those areas of southern Arabian Sea which have both ASHSW and RSW. An important feature of the vertical salinity structure in the Arabian Sea is the transient multiple subsurface maxima separated by pockets of low saline water. A subsurface salinity minimum is found at about 150 m depth and its aerial extent is highest during November-January. The factors leading to its formation appears to be (i) transport of low salinity water by the northeast monsoon coastal current along the west coast of India and subsequent increase in the surface salinity of the water due to evaporation in the north Arabian Sea and (ii) poleward coastal undercurrent along the west coast of India during the southwest monsoon.

An influx of fresh water of about 28.1×10^3 m^3 s^{-1} flows in to the Bay of Bengal annually and about 50% of this comes during the summer monsoon months. During the summer monsoon, low salinity water (~29.0‰) spreads into the interior of the Bay of Bengal in a southwesterly direction from the Head of the Bay and the northern Andaman Sea. At 100 m depth, during the pre-summer monsoon, the salinity distribution is characterized by a low salinity cell with a core value of 36.6‰. During the summer monsoon a tongue of high salinity with a core value of 35.2‰ protrudes into the central Bay. This high salinity water is seen due to the penetration of ASHSW into the central Bay from the southwestern Bay at subsurface depths (75-100 m). It has a uniform value of salinity (34.9‰) at 100 m depth in the northwestern Bay. The ASHSW is not found in the southern Bay during the pre-monsoon period. At 200 m depth, during the pre-summer monsoon salinity variation is marginal (34.90 – 35.05 ‰). During the summer monsoon season, salinities all over the Bay are over 35.0 ‰ indicating the stronger intuition of Indian Equatorial Intermediate Water. During the winter monsoon, a large cell of low salinity (34.90 ‰) occupies the central Bay. At 500 m depth during the summer monsoon the salinity distribution is characterized by zonally oriented isohalines with relatively high salinities (> 35.06‰) in the central Bay and southern Bay and low salinities (< 35.06‰) north of 16° N. The penetration of subsurface high salinity layer is maximal during summer and post-summer periods. The whole Bay of Bengal is occupied by this high saline water during summer and winter. At 1000 m depth during the pre-summer monsoon a cell of low salinity (34.94‰) appears in the central Bay. During the summer season, lower salinities (34.92 ‰) in the northwestern Bay and higher salinities (34.98‰) off the southern east coast of India are seen. During the post-summer season, salinities are much less compared to the summer season with the lowest salinities (34.88 ‰) near the Nicobar Islands.

During the winter monsoon, along the 91° E meridional section two pockets of low salinity centered at 8° N and 15° 30' N are seen in the upper 50 m. The salinity of the northern pocket is comparatively low (31.4‰) due to the influence of fresh water influx from the Irrawady river. The depth of the halocline is 100 m at 15° 30' N where as it is about 75 m between 7° 30' N and 9° 30' N. The intermediate high salinity water exhibits large depth and thickness variations from south to north. Along 50° N zonal section surface salinity decreases from 33.8‰ in the west to 31.6 ‰ in the east. In the upper 100 m, in the western part of the section relatively high salinities are observed in association with the northward flow. During the summer monsoon along 89° E meridional section low salinity (< 33.8‰) waters are seen north of 12° 30' N with salinity decreasing northward to 32.6 ‰. Two cells of high salinities (35.0 – 35.2 ‰) identified as the result of penetration of ASHSW are located around 50 m at 12° N and around 40 m at 14° N. The intermediate high salinity layer is seen with its core value of more than 35.1‰ at 8° N. At depth

greater than 500 m, relatively low salinity waters are seen north of 12° N. along 13° N zonal section, high surface salinities (33.6‰) are seen off the Chennai coast. Low salinities (32.4 ‰) are observed at 91° 30' E because of the influx of the fresh water from the Irrawady River. The halocline is located at about 75 m. High salinity cells appear at depths of about 90 m due to penetration of high salinity water mass. The intermediate high salinity layer is divided into large cells occupying the western and eastern regions. The meridional distribution of salinity along 95° E in the Andaman Sea during the winter monsoon shows low surface salinities (32.0‰) owing to the runoff from the Irrawady River.

The thickness of the Intermediate High Salinity water is about 350 m and the core salinity is slightly less in the northern Andaman Sea. Salinity, in general, increases with depth to give a salinity maximum at the top of thermocline with in the equatorial waters. This subsurface salinity maximum is generally found in the north equatorial Indian Ocean during the northeast monsoon. The presence of subsurface salinity maximum with salinity greater than 35.1‰ between 500 m and 900 m at 2° to 4° N in the western equatorial Indian Ocean is attributed to the penetration of Red Sea Water. Near 10° S, the hydrological structures of northern and southern parts of the Indian Ocean are separated by a hydrological front marked by a salinity minimum found to a depth of about 500 m. The low salinity waters (< 34.70 ‰) occupying the upper 100 m between 9° and 14° S along 79° E section are brought in from the east by the south equatorial current. A subsurface salinity maximum was also located around 250 m depth. The presence of a very strong halocline within the upper 20 m around 70° E along 10° S section is attributed to the surface dilution by heavy precipitation associated with the Inter Tropical Convergence Zone.

5.8.4. Barrier layer (BL):

The layer between the halocline and the thermocline is referred to as the Barrier Layer (BL), because of its effect on the mixed layer heat budget.

A BL is present in the Bay of Bengal through most of the year due to advection of low salinity water diluted from monsoon river runoff. The fresh water influx into the Bay of Bengal maintains a BL, which has a thickness of 25 m during the summer monsoon. From time series of hydrographic observations conducted in the northern Bay of Bengal at 20° N, 89° E during August-September 1990 (MONTBLEX), found that the mixed layer based on a temperature criterion is deeper than that using density. The formation of a BL was observed in the Bay of Bengal during summer monsoon (BOBMEX) from time series measurements of temperature and salinity profiles in the upper ocean at 17.5° N, 89° E from 27 July to 6 August 1999. Initially the upper layer that is homogeneous in both temperature and salinity was about 30 m deep. Subsequently, the arrival of a fresh water plume (from river discharge and rainfall) caused MLD to decrease to about 10 m and salinity in the surface layer by about 4‰. The plume led to the formation of a new halocline and hence a BL (of about 15 m thick) within the upper 30 m of the water column. The ensuing ocean-atmosphere interaction was restricted to the new thinner mixed layer. The cooling that was restricted to the mixed layer led to an inversion in temperature (0.5 C) just below the mixed layer.

The barrier layer is the layer within the isothermal layer and below the mixed layer within which the salinity increased rapidly (blue stippled region in Fig.5.45). Barrier layer was numerically calculated by subtracting the MLD from the isothermal layer.

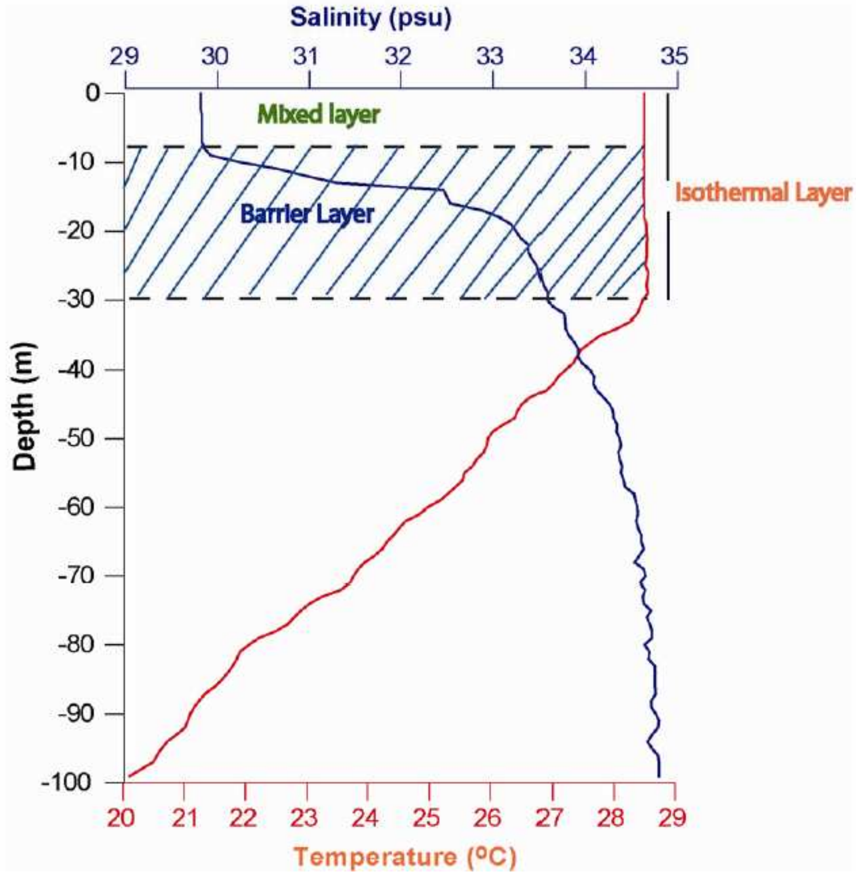


Fig.5.45. Vertical profiles of temperature and salinity in the northern Bay of Bengal (19^{O}N, 88^{O}E) during August 2001 depicting the isothermal and barrier layer (By courtesy of Narvekar 2008)

From time series measurements of temperature and salinity profiles conducted in the southeastern Arabian Sea during 27 March-7 April and 23 May-7 June 2003 as part of ARMEX, it was found that a 20 m thick BL exists during March-April owing to a surface layer of low salinity waters advected earlier during December-January from the Bay of Bengal. The BL is annihilated during April-May by upwelling and inflow of high salinity waters from the north and by mixing due to stronger winds, which deepen the mixed layer.

5.8.5. Barrier Layer Thickness (BLT):

The barrier layer thickness (BLT) showed large spatial variability during January, varying from 70 m in the northern Bay and eastern equatorial region to less than 30 m in the central and southern Bay (Fig.5.46a). The region around Sri Lanka and peninsular India had a deep BLT. The basin-wide BLT was the least in April, which was about 20 m, except close to the eastern part of the equator where it was about 30 m (Fig.5.46d).

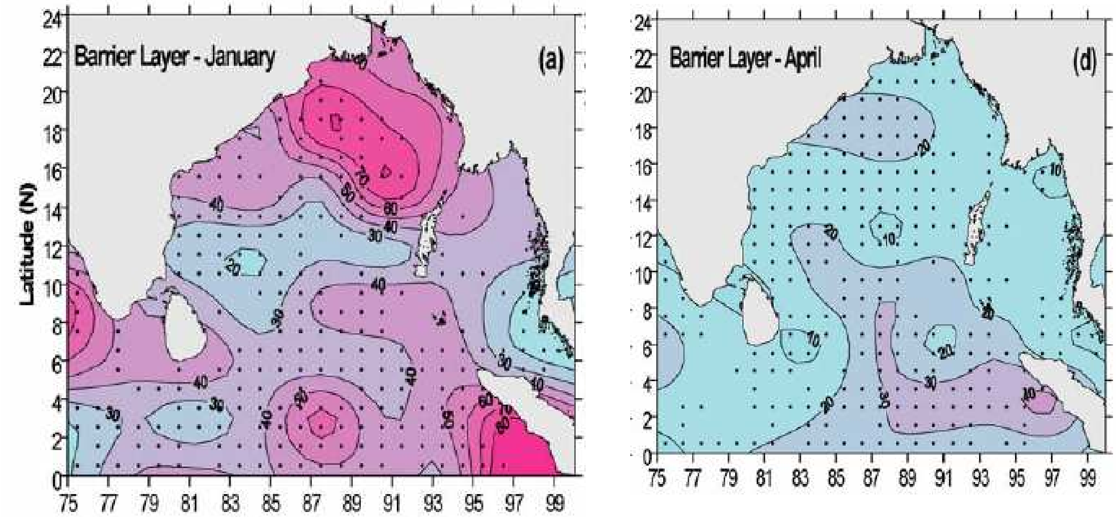
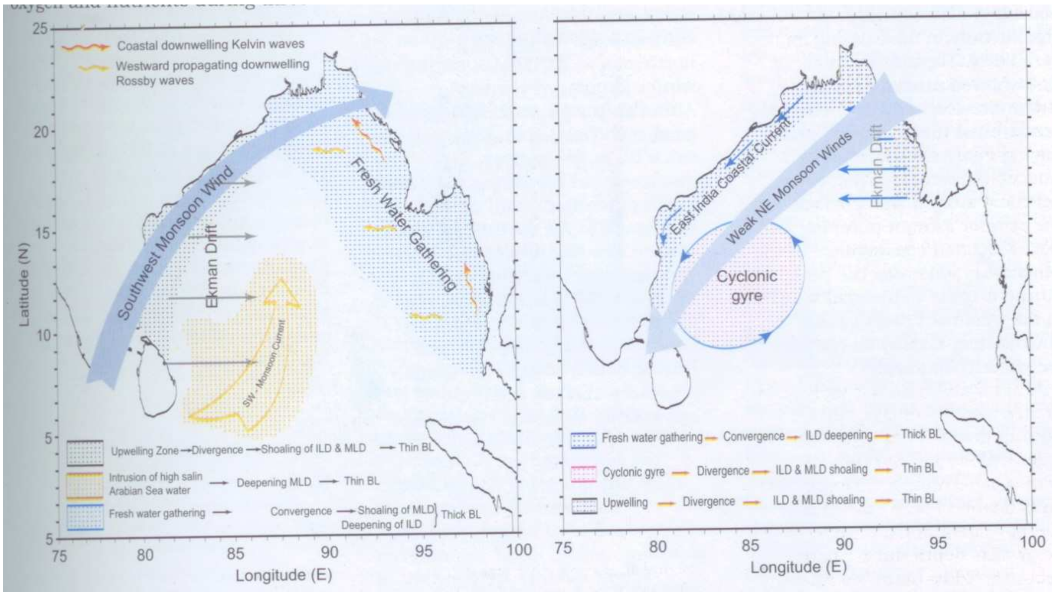


Fig. 5.46 The variation of thickness of barrier layer in January(a) and April(d) in bay of Bengal (Narvekar 2008).

Seasonal variability of barrier layer in the Bay of Bengal:

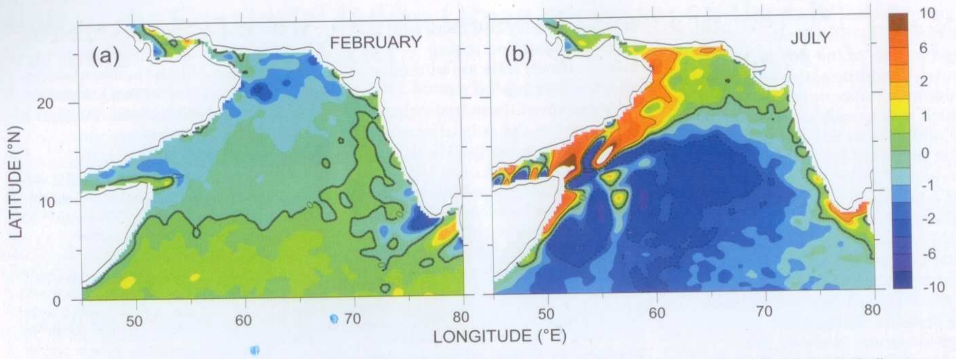
The formation of barrier layer (BL) and the seasonal variability of barrier layer thickness (BLT) in the Bay of Bengal were examined utilizing the most comprehensive data set. Large BLT of 40 m first appeared in the coastal regions of the northeastern Bay in June and spread westward as the summer monsoon progressed. Along the east coast of India the formation and variability of BL was controlled by East India Coastal Current (EICC). Another high BLT of 50 m appeared along the east coast of India in November when the EICC flow was equatorward and was spatially organized by December. By January it weakened when the EICC reversed. The mature phase of BLT, both in amplitude (60 m) and spatial extent occurred during February, when the subtropical anticyclonic gyre was well established in the Bay of Bengal. During both the summer and winter monsoon seasons the surface circulation and the redistribution of low saline waters showed a dominant influence on the observed BLT distribution. Other processes such as Ekman Pumping and propagating Rossby waves forced by the propagating Kelvin waves along the eastern boundary also contributed significantly in modulating its variability. The annual mode of BLT showed maxima during November-December, whereas the semi annual mode showed peaks during February-March and August-September. The peak in Feb-March was attributed to the interior Ekman pumping and the associated convergence in the central bay, whereas the peak during August-Sept was due to the westward propagating down welling Rossby waves from the eastern boundary.



Schematic of the forcing mechanisms explaining the variability of ILD, MLD and BLT for the summer monsoon (a) June–September (b) November–December

Seasonal variability of barrier layer in the Arabian Sea:

The formation and mechanisms of the BL and its seasonal variability in the Arabian Sea (AS) are studied using a comprehensive dataset of temperature and salinity profiles from Argo and other archives for the Arabian Sea. Relatively thick barrier layer of 20-60 m with large spatial extent is found in the central south western Arabian Sea, the convergence zone of the monsoon wind, during the peak summer monsoon (July-August) and in the southeastern Arabian Sea (SEAS) and northeastern Arabian Sea during the winter (Jan-Feb). Although the BL in the SEAS has been reported before, the observed thick BL in the central southwestern Arabian Sea (CSWAS) during the peak summer monsoon and in the northeastern Arabian Sea (NEAS) during later winter are the new findings. The seasonal variability of BLT is closely related to the processes that occur during summer and winter monsoons. During both the seasons, the Ekman processes and the distribution of low salinity waters in the surface layer show a dominant influence on the observed BLT distributions. In addition, Kelvin and Rossby waves also modulate the observed BLT in the AS. The relatively low salinity surface water overlying the Arabian Sea high salinity water provides an ideal ground for strong haline stratification in the CSWAS during summer monsoon and in NEAS during winter monsoon. During summer, northward advection of equatorial low salinity water by the Somali current and the offshore advection of low salinity water from the upwelling region facilitate the salinity stratification that is necessary to develop the observed BL in the CSWAS. In the SEAS, during winter, the winter monsoon current carries less saline water over relatively high salinity ambient water to form the observed BL there. The winter west India coastal current transports the low salinity water from the SEAS to the NEAS, where it lies over the subducted ASHSW leading to strong haline stratification. Ekman pumping together with the downwelling Kelvin wave in the NEAS deepen the thermocline to cause the observed thick BL in the NEAS.



Wind stress curl (10^7 N m^{-3}) for (a) February and (b) July computed using the QuikSCAT winds during July 1999–May 2006. Shading interval is 1.0 between [10, 10]

5.8.4. Salinity along the Coastal Regions of Bay of Bengal:

The average surface salinities were found to vary from 33.27 to 33.81‰ during October 1921 to February 1922 in Nankauri harbor (Nicobar Islands). It is known that large quantities of fresh water enter the Bay of Bengal from the rivers Ganges and the Irrawady and cause in the reduction of surface salinity in the northern Bay of Bengal. This is the reason for the low salinities of less than 34‰ during summer monsoon in the greater part of the surface of Bay of Bengal. But in Palk Bay salinities between 34.5 and 35.0‰ were observed. Surface salinities of about 29‰ in January, 34‰ in March and less than 18‰ during summer monsoon off Waltair were found. Bottom salinities were found to be between 33.5‰ to 34.5‰ over the central shelf. These vertical salinity gradients reflect about the runoff and circulation. It was also reported high surface temperatures during July-October, when flow of fresh water into the Bay is maximum and low when the discharge is minimum (December-February).

Some workers reported that with increased run off during summer monsoon salinity values off the river mouths in the northern Bay decreased to about 18‰ and in southern area to about 34‰ through 30‰ in middle sector. Low salinity (<20‰) values were observed near the mouths of river Irrawady during summer monsoon. Deepening of halocline and low salinity waters off the mouths of Krishna and Godavari were also noticed. The water column over the continental shelf off Thailand was observed to be strongly stratified during winter monsoon and weakly stratified during summer monsoon.

During the winter monsoon the surface salinities range between 27 and 33‰. The 33‰ isohaline runs almost parallel to the east coast of India. Lowest salinity (27‰) is observed in the region around 20°N due to the Ganges- Brahmaputra system. At 50m the 33‰ isohaline breaks at different regions and high saline waters reach near shore areas. At 100m in most of the shelf the salinities vary between 34.6 and 34.9‰ except at river mouths. At 200m salinity is more uniform with insignificant mixing with top diluted water (34.5‰). At 500m it varies from 35.05 to 35.10. During winter monsoon along 14°N (Off Chennai) surface salinities vary from 32 to 33.4‰ from about 100 to 300 km offshore compared with 27 to 32.5 along 20°N (off Mahanadi) from about 50 to 300 km offshore. The diluted water (<34.5‰) is observed only up to 40m depth. Along this section the salinity contours tilt upwards offshore to the surface. Off the central coast at about 17°N surface salinities vary between 27‰ near the shore to 33.5‰ at about 200 km. diluted water extends up to 200 km and vertically occupies about 60 m of the water column. During the same season in the northern region (20°N) surface salinities vary from 27‰ near the shore to 32.5‰ near off shore (300 km). The diluted water (<34.5‰) extends up to 300km in the horizontal and up to 50 km in the vertical. Surface salinities are greater than 30‰ off the Ganges for about 200 km in the horizontal and about 75m in the vertical. The surface salinity increases from 32‰ to

33‰ from north to south in the Andaman Sea in winter. Surface salinity variations in the coastal waters off the west coast of Thailand were found to be 32.1‰ and 33.6‰ during Jan-Feb 1983. The diluted water (<34.5‰) occupies the entire shelf over the northern and eastern areas and extends up to 75 m in the vertical in the full northeastern Bay.

During the summer monsoon (Fig.5.48) surface salinities range between 22 to 34‰ over the east coast of India. The 33‰ isohaline forms a boundary across the central east coast (about 16°N) between the southern high salinity and northern low salinity belts. Off Chennai the diluted water extends up to 50 m with more or less same thickness. Off Viskhapatnam (17°N) the diluted water extends up to 70 to 90 m with a wavy structure. In the northern most section the diluted water extends upto > 110 m with almost uniform thickness.

The range of variation of surface salinities from summer (12‰) to winter (6‰) is twice. The surface diluted water shows wavy distribution across the shelf irrespective of the season. The fluctuations of ranges in the coastal areas (22-34‰) are more than that in the deep ocean (29-34.5‰) in summer and also similar in winter.

The low salinity tongue extending down to Visakhapatnam from north is caused due to southward flowing coastal current along east coast of India (East India Coastal Current). The high salinities in the southern region are due to the northward extension of the Indian Monsoon current.

During winter (Fig.5.47), the relatively low salinity waters flow into the Bay of Bengal through the westward flowing North Equatorial Current which mixes with the clockwise gyre in the southern Bay of Bengal. This seems to be the reason for the continued low salinity coastal belt along east coast of India.

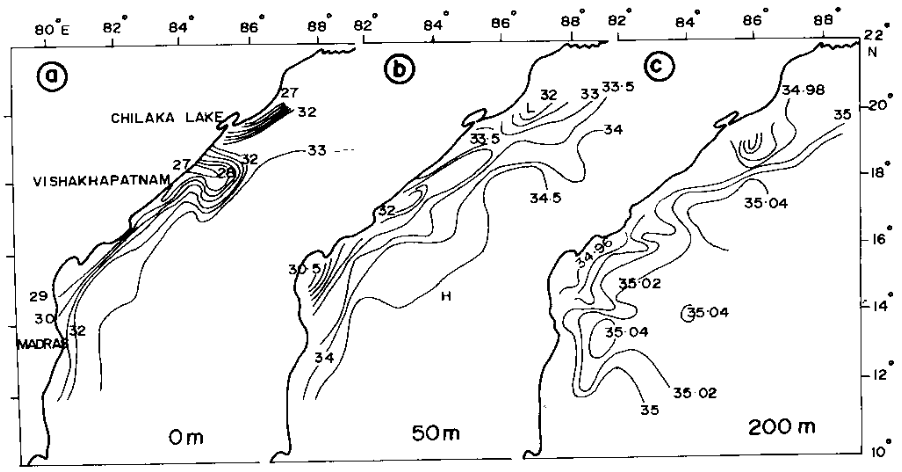


Fig.5. 47. Distribution of salinity during winter at various levels in coastal areas of Bay of Bengal. (By courtesy of Suryanarayana et al.,1992)

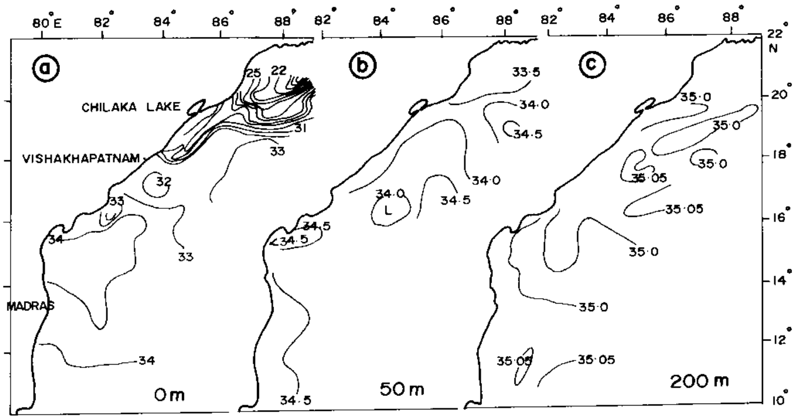


Fig.5.48. Distribution of salinity during summer at various levels in coastal areas of Bay of Bengal. (By courtesy of Suryanarayana et al., 1992)